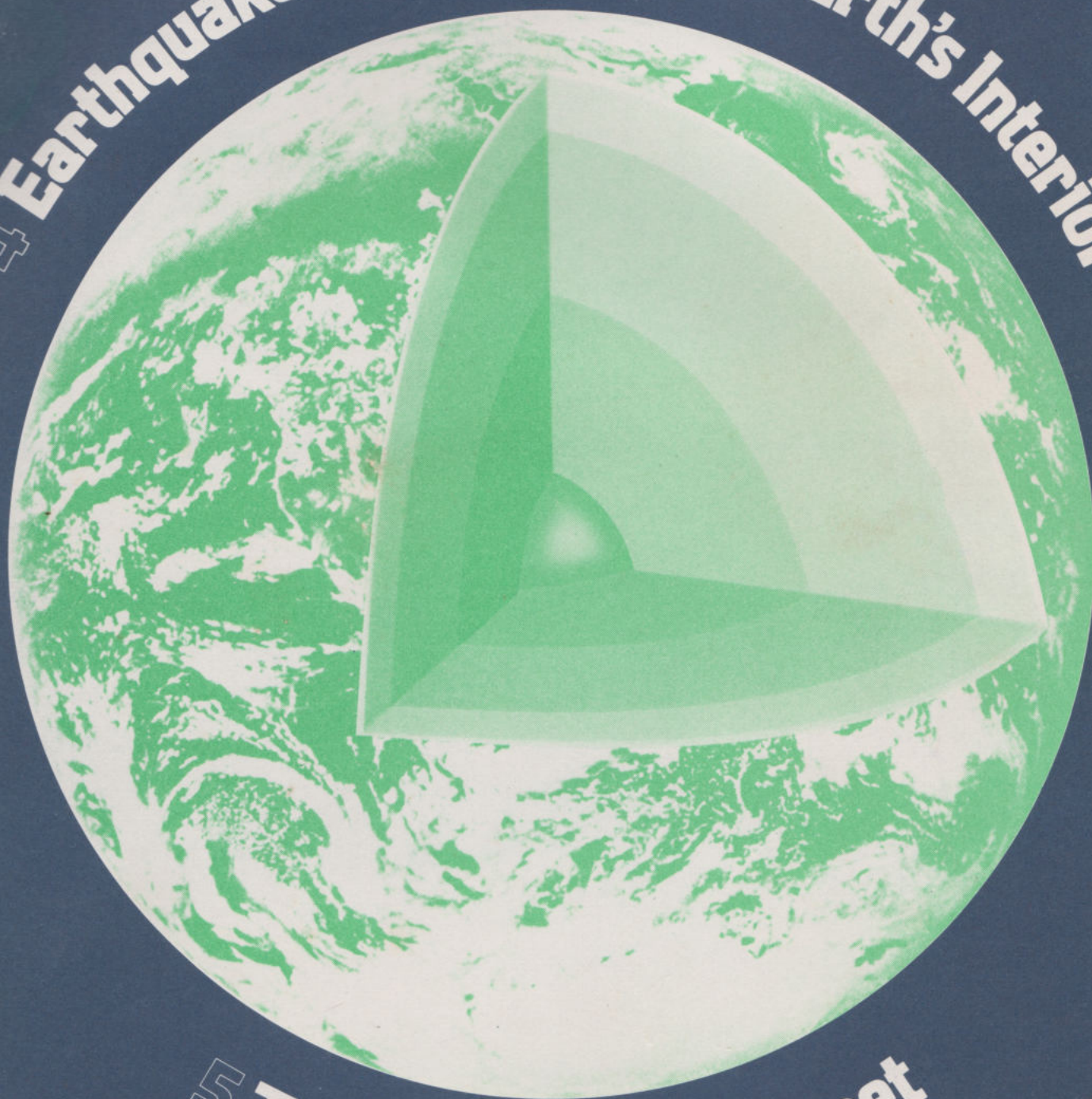




4 Earthquake Waves and the Earth's Interior



5 The Earth as a Magnet



The Open University
Science: A Foundation Course

Unit 4

Earthquake waves and the Earth's interior

Prepared by the Science Foundation Course Team

The Open University Press

SCIENCE

S101 Course Team List

A note about the authorship of this text.

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TABLE A List of terms and concepts in Unit 4

Assumed from general knowledge	Defined in a previous Unit	Unit No.	Introduced or developed in this Unit	Page No.	Introduced or developed in this Unit	Page No.
cross-section	acceleration	3	angle of incidence	26	magma	31
earthquake	conceptual model	1	angle of refraction	26	mesosphere	46
elastic medium	constant of proportionality	MAFS	asthenosphere	46	Mohorovičić discontinuity	43
erosion	density	2	axial modulus	19	ocean crust	44
fault	dimensionless ratio	MAFS	basalt	31	outer core	42
friction	gravitational force	3	body waves	34	P-waves	11
meteorology	inertia	3	compression pulse	9	partial melting	46
microscope	mass	3	compression wave	9	peridotite	31
mineral	newton (N)	3	compressional strain	18	phase changes	45
nuclear weapon	Newton's laws of motion	2	compressional stress	18	refraction	24
sound	planet	1	continental crust	44	rigidity modulus	21
surface waves on water	solar system	1	continuous refraction	29	S-waves	11
telescope	velocity	2	core-mantle boundary	38	sandstone	31
volcano			critical angle	28	sedimentary rock	31
volume			earthquake epicentre	15	seismic discontinuity	24
			earthquake focus	15	seismic zone	15
			earthquake magnitude	16	seismogram	17
			earthquake intensity	16	seismology	6
			effusive volcanism	31	seismometer	7
			elastic modulus	18	shadow zone	38
			energy	11	shearing force	13
			epicentral angle	33	shearing strain	21
			explosive volcanism	31	shearing stress	21
			extrusive rock	31	stress	18
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Introductory study comment

Unit 4 brings you from the realms of measurement and the solar system right down to the Earth beneath our feet. What do we know about the interior of the Earth and how do we find out? The title of this Unit might suggest an answer to you! You will be studying first the properties of earthquake waves and their passage through the Earth, and will follow this with deductions about the physical properties of the Earth's interior that will lead to a model for its composition.

The study of the Earth, both its surface and deep interior, comprises the science of geology or, more literally, Earth science. Rocks are the fundamental materials and, in the first Earth science Unit, we introduce you to the rock samples in your Home Experiment Kit. Right at the outset (Section 1.2), we want you to measure the densities of some rocks and, later, this will help you to understand at what depth they occur most commonly inside the Earth. The experiment will take about 1½ hours and *must* be completed before you start Section 4.

Later on, at the beginning of Section 4, you will need an hour to study the Audio-vision sequence associated with this Unit, which describes the processes of rock formation. Again, this is designed to aid your understanding of this and future Earth science Units: it is one of the most important parts of Unit 4.

1 Seismic waves and their use in determining the internal structure of the Earth

1.1 A study of the inaccessible

In Units 1 and 2 you studied the Earth as a planet in the solar system, and in the process you were introduced to a *way of thinking* about objects or phenomena in nature that is important and typical of science. This is the idea of building a hypothetical *model* of a system so that you can understand it and hence make predictions about its behaviour that you can test experimentally.

Although we can often extend the range of our senses with the help of instruments such as telescopes, microscopes and satellites, most of the objects and phenomena in the Universe are not accessible to us for *direct* observation. Beyond the limits of the Earth and the Moon, we reach the inaccessible, where our understanding of nature is expressed in terms of models that may combine diagrammatic, verbal or mathematical descriptions. For example, when you describe qualitatively how the Earth rotates on its own axis and goes round the Sun in an elliptical orbit, you are using a *verbal model*. To provide a better understanding of the concept you could use a *diagrammatic model*, as indeed did many of the Figures you studied in earlier Units. A *mathematical model* could be used to quantify the Earth's orbital period, as in the inverse square law of gravitational attraction (Unit 3, Section 5.3). Often, by combining the attributes of different modelling techniques, we can obtain a much better understanding of the phenomenon being studied.

In view of our strong dependence on the Earth, the planet on which we live, perhaps it is somewhat ironic that the Earth's deep interior is even less accessible to direct observation than outer space. To understand the nature and composition of the Earth we must resort to indirect techniques, which allow us to build up and refine a descriptive model. This study will enable you to appreciate how scientific models can be used to study the inaccessible.

To appreciate just how inaccessible to direct observation is the Earth's interior, consider the following: detailed exploration of the Earth's surface is well advanced, and man has mined for metals 'deep within the Earth' and tapped reserves of mineral oil from 'deep' boreholes. Yet these workings penetrate at most a few kilometres. Compare this with the 6370 km radius of the Earth! Imagine a scale model of the Earth in the form of a globe 50 cm in diameter—the size of a large plastic beach ball. Even man's deepest workings, on this scale, would not penetrate the thin plastic skin and let the air out! Geological processes have brought rocks to the surface that were originally created at depths of up to about 25 km; on our model, even this depth is still within the beach ball's plastic skin. Volcanoes erupt molten rock originating some 60–100 km down within the Earth. Although this molten material has probably been grossly altered on its way to the surface, it nevertheless provides a small amount of evidence about the composition of the part of the Earth from which it came—but even 100 km depth is only just inside the surface.

Without being able to make direct observations, how can we obtain any information about the composition and structure of the Earth's deep interior?

1.2 The Earth's shape and average density

Before you proceed any further with your study of the Earth, we should like you to complete the Home Experiment for this Unit. This is because you will measure the density of some common rock samples, contained in the Home Experiment Kit, and then compare their densities with that of the whole Earth. In this way, you can make meaningful deductions about the Earth's interior and how similar it might be to your rock samples. You should refer to the *Home Experiment Notes* for this Unit for full instructions, and complete the experiment now if at all possible, but certainly before you reach Section 4. When you have completed the experiment, answer the following question.

ITQ 1 In Section 6 of Unit 3, you calculated the average density of the Earth. Recall the value you obtained and compare this with the densities of the four rock samples from the Earth's surface which you have just determined. Does this comparison lead you to any conclusions about the Earth's interior?

So, merely by 'weighing the Earth', it is possible to find out *something* about its deep interior.

The *shape* of the Earth also tells us something about the interior. It turns out that the Earth is not quite spherical, it has an elliptical cross-section, slightly flattened at the poles. Very accurate measurements of the Earth's shape have been made possible by remote satellite observations and by sensitive gravity meters, which show that the force of gravitational attraction changes over the Earth's surface. We call such a flattened sphere-shaped body as the Earth an *oblate spheroid* (Figure 1).

Recalling from Section 5.3 of Unit 3 that the gravitational attractive force for a mass (m) at the surface of the Earth (radius r , mass M) is given by $F = GMm/r^2$, will the downwards force exerted by gravity be greater at the equator or the poles?

Since r is smaller at the poles, $1/r^2$ must be greater, so that, assuming G , M and m are constant, F will be greater at the poles than the equator. (We have assumed that, in calculating the force at the surface, the mass of the Earth may be considered to be located entirely at the Earth's centre.)

There are other effects that cause variations in gravity over the Earth's surface, but these need not worry us at this stage.

Sir Isaac Newton knew that the Earth was flattened in this way and he suggested, quite rightly, that the equatorial bulge was due to rotation, with a concentration of matter at right angles to the Earth's rotational axis. This means that parts of the interior can deform under stress and the subsequent discovery of liquid layers, such as the outer core, was not unexpected.

Like the Earth's average density, shape tells us only a very little about our planet's interior. As you will see in Unit 5, the fact that the Earth has a 'magnetic field' (which man has used for ages as an aid to navigation), and the way this magnetic field has changed from time to time, does provide some more indirect evidence about the gross structure and composition of the Earth. Volcanoes also provide a clue—they tell us, in quite spectacular fashion sometimes, that the interior of the Earth must be very hot.

Although you are now in a position to recognize that the density of the Earth increases towards the centre, you have, as yet, no means of telling whether there is a smooth increase from surface to centre, a series of jumps in density, or any other distribution you might care to suggest. Yet, as you will realize from the Home Experiment Notes, density is a most useful physical quantity, because it characterizes the composition of a material—in this case, the Earth's interior. Important as it may be, the information from volcanoes and the magnetic field are not going to help very much in deciding between alternative density distributions.

Can you think of any other natural geological phenomenon that might prove a more useful means of probing the Earth's interior?

The most useful evidence comes from the study of earthquakes and the shock waves they generate, and this is because they travel right *through* the Earth.

The study of earthquakes is called *seismology*, and the velocity with which the waves travel tells us something about the physical characteristics, including the density, of the material through which they pass. For this reason, much of the remainder of this Unit is devoted to seismology.

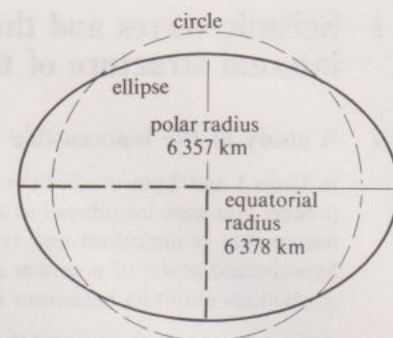


FIGURE 1 Cross-section through the Earth.

seismology

1.3 Earthquakes

Earthquakes occur much more frequently in some places, like the Middle East, Japan and California, than they do in others—by the time you have completed Unit 7 you will understand *why* this is so. They can often produce violent movements of the surface and can cause severe damage to structures near the centre of a disturbance. A large earthquake will cause damage tens, hundreds and occasionally thousands of kilometres away from its point of origin. The same earth-

quake will produce a slight tremor at points all over the Earth's surface, too weak to feel, but easily detectable by instruments designed for this purpose and called *seismometers* (see Section 2.4). When an earthquake occurs, the disturbance produced radiates out in all directions as indicated in Figure 2. Some of the disturbance travels along the surface layers of the Earth, but most travels *through the interior of the Earth*. For example, earthquakes in New Zealand can be detected on the opposite side of the globe in Europe, and we know from the type of wave motions detected (Section 1.4.4) that some of the disturbance has travelled right *through the deep interior* of the Earth. The same applies to any earthquake: it can be detected over almost the entire surface of the Earth.

seismometer

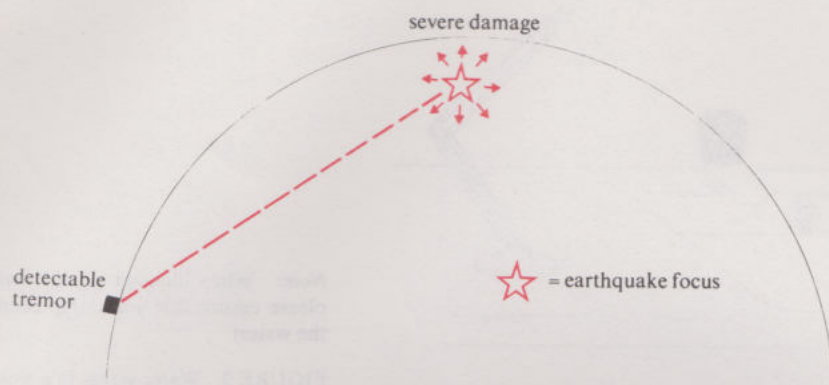


FIGURE 2 Earthquake disturbances travel out in all directions from the source.

Fortunately—although perhaps only for those interested in the deep Earth—earthquakes are very frequent: about 150 000 are recorded every year. In recent years, seismometers have become increasingly sensitive and thereby capable of detecting extremely small tremors. In fact, they are sensitive enough to detect the effects of an underground explosion of a small nuclear weapon many thousands of kilometres away and, what is even more important, to distinguish it from a natural earthquake. The military and political importance of being able to do this is one reason why there are many more sensitive seismometers around the world than thirty years ago.

How can these *earthquake waves* be used to provide specific information about the materials through which they have travelled? In Section 2 of this Unit you will be concerned with where and how earthquake waves are produced and detected; in Section 3, with how they are propagated; and in Section 4, with how they can be used to find out about the structure and composition of the Earth's interior. But before we go any further, why are we talking about waves at all?

1.4 Earthquake waves: what are they and how do they travel?

You may think we are using terms like 'earthquake waves' and 'shock waves' without being at all specific about what we mean by *waves* or about why we have this notion of earthquake waves radiating out from a point inside the Earth where a rock fracture occurs. What do we mean by 'waves'?

1.4.1 Water surface waves

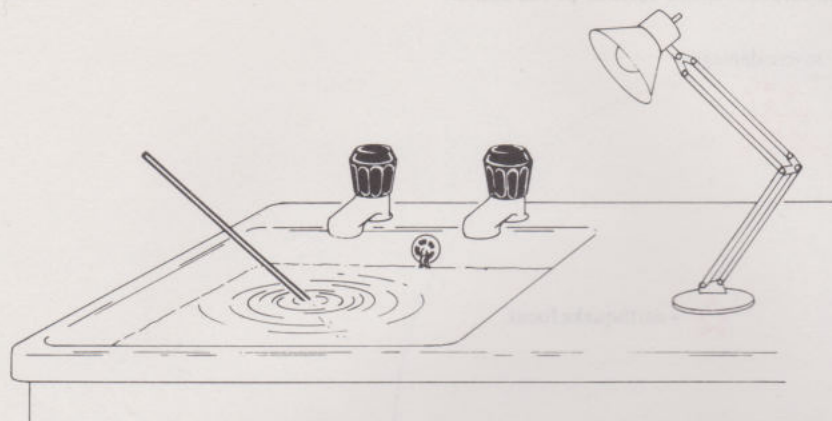
Probably the most familiar example of waves is that of surface waves on water. No doubt at some time you have tossed a stone into the middle of a pool and watched the ripples spread out across the quiet surface. Next time you are near a pond, why not repeat this simple 'experiment', but this time with the sharp eye of a science student! Toss a few small twigs or bits of cork onto the surface before you drop the stone in the middle, and watch what happens to them as the ripples pass them by. Ask yourself: what is moving across the surface of the pond?

Try another simple experiment in your kitchen sink or in the bath. All you need is a pencil and a small piece of cork or similar buoyant substance, but you will see more if you illuminate the water surface with a bright light (a spotlight or reading

lamp) at an oblique angle, as in Figure 3. Then, with the tip of the pencil, touch the water surface and keep on touching the same spot regularly at about one second intervals. What happens? Something travels across the surface of the water. What is it that travels: is it the water itself? How can you check this?

If the water travels, it will carry the cork with it. Drop the cork in the water and try again. The cork on the surface bobs up and down but does not move outwards. It is the disturbance caused by the pencil that travels out across the surface. The transmission of the disturbance is called *wave propagation*.

wave propagation



Note: When illuminating the water surface, please ensure that your lamp cannot fall into the water!

FIGURE 3 Water waves in a sink.

1.4.2 Waves in solids

Now try another experiment; this time you may need a helper. Also, you will not be able to see the waves being propagated, but you will be able to make some controlled guesses about what is going on. First you need a table—the longer the better. Sit at one end of the table with one ear pressed against the top surface. Block the other ear off with some cotton wool. Now, get your collaborator to tap the table *very gently* at the opposite end, with a pencil or any convenient object of that sort. While the tapping continues, lift your ear off the table top—a few inches will do—and then press it back again. Try it a few times and note your impressions. Next try quietly scratching the table instead of tapping it. What do you notice this time—again make a few notes.

Did you detect the disturbances more clearly when your ear was on the table or off it? Obviously, when your ear is on the table, you are detecting the disturbance as it moves through the wood of the table; when your ear is away from the surface, you are detecting the disturbance that travels through the air. Evidently, the disturbance produced at one end of the table is propagated much better through the material of the table than through the air above it. A water pipe between two rooms of a house shows this effect very well. It is, of course, a well-known way of signalling used by prisoners.

But again, *what* travels? It is obviously not the table, or any part of it, that travels. It is simply 'the disturbance', a state of vibration, that is transmitted. When the disturbance reaches the end of the table under your ear, the vibrations are transmitted from the table to the air, and thence to your ear as sound. But is this vibration *all* that is transmitted? To help you think this one out, imagine a row of balls, connected by springs and suspended by very fine threads, as shown in Figure 4. We shall show you a model like this in the television programme associated with this Unit (TV 04), but you should be able to do this experiment 'in your mind's eye'—a *thought experiment*.

'thought experiment'

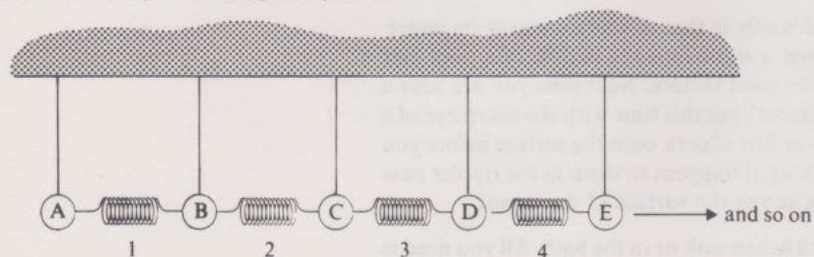
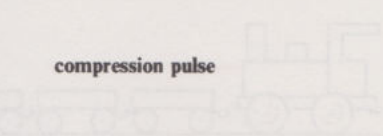


FIGURE 4 Ball-and-spring model for wave propagation.

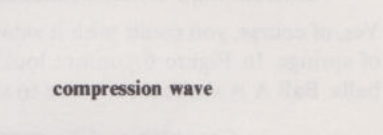
Suppose you were to give ball A in Figure 4 a sharp push to the right. What would happen?

As ball A moves towards ball B, it compresses spring 1. But the spring is elastic and does not stay compressed; it returns to its original length and thereby pushes ball B to the right and ball A to the left. Ball B, which in turn moves to the right and back again, squeezes spring 2 and transmits the same effect to ball C. This process continues down the line of balls and springs: it represents in our model (Figure 4) the travelling of a *pulse*. Thus a compression pulse travels down the line of springs and balls. Apart from that, because the springs are elastic, they will come back eventually to their original lengths (though they may 'overshoot' a few times in the process), and on each return towards the left an expansion of rarefaction pulse is observed. So a *compression pulse* has passed through the medium—down the line of springs—and after they have returned to their original lengths the medium is left *exactly as it was before*. Each ball oscillates about its 'mean' position, at which it eventually comes to rest when the disturbance has passed. Note that, in this thought experiment, each time a ball moves it is experiencing a *force* that has been transmitted from the original motion of ball A via the springs. The amount and rate of movement caused by this force, indeed the size of the force on each ball, depends on how stiff the springs are, as we shall see shortly.



compression pulse

Now imagine that instead of giving ball A a single jerk to the right, you set up some machine to vibrate A in the plane of the paper from left to right (away from and towards ball B) in a regular rhythm. What would happen? You would see a series or *train* of compression pulses travelling down the line one after another. Think again about the water waves travelling across the surface of the pond, basin or bath. What you actually *see* is a series of peaks and troughs, where the water is displaced up and down, moving, one after another, across the surface. With the line of balls and springs, what you see is a series of places down the line where the springs are compressed and, in between, a series of places where they are extended. These compressions travel down the line and so we use the words *compression wave* by analogy with the waves with which we are most familiar: water waves.



compression wave

How do we apply our thought experiment to the solid materials of the Earth? On a sub-microscopic scale, we can imagine rocks to be three-dimensional arrays of balls and springs and, although it would be difficult to *see* the disturbance set up by compression waves in incompressible rock, such waves actually are transmitted (remember your experiment with the table!). The rate at which compression pulses are transmitted depends on certain properties of the 'network of balls and springs'. Consequently, if we can determine the speed of the earthquake wave compression pulses in the Earth, we should be able to determine certain properties of the rock media involved.

1.4.3 Wave velocities—what *should* they depend on?

For simplicity, think again about the row of balls and springs shown in Figure 4 and ask yourself the following questions:

Would the compression pulse travel faster or slower down the line of balls and springs, (a) if the balls all had greater mass, but the springs were unchanged; (b) if the springs were stronger, i.e. needed more force to compress or elongate them by a given amount, but the balls were unchanged and were displaced by the same amount?

Recall the relationship between force, mass and acceleration given by Newton's second law of motion (Unit 3, Section 3.6): $F = ma$. If we increase m , there will be greater inertia, and so a given force F will produce a smaller acceleration a , and hence a smaller initial velocity. So, in Figure 4, if the balls were made *more massive*, a given force will cause the pulse to travel *more slowly*.

To answer question (b), bear in mind that *stronger*, stiffer springs mean that *more force* will be exerted by one ball on its neighbour for a given amount of spring displacement. In terms of Figure 4, each spring, as well as transmitting greater forces down the line, will also take less time to compress if the stiffness is increased. This means that each ball will be set in motion *more quickly*, and we see also from $F = ma$ that, for a given mass, the greater the force F , the greater the acceleration a , and hence the greater the resulting velocity. So, a *stiffer, stronger* set of *springs* will cause the pulse to travel *more quickly*.

Here is another 'thought experiment' which might help you see these points more clearly. Imagine a series of firmly coupled railway trucks parked in a siding (Figure 5). A shunting engine comes up and gives the end truck a nudge, displacing it a few inches to the right. The couplings (and the trucks) being pretty incompressible, the train strongly resists being compressed, and so the push is quickly transferred along the train and the truck at the far end moves a few inches to the right.

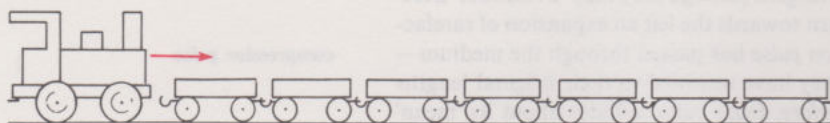


FIGURE 5 Firmly coupled set of railway trucks.

ITQ 2 Now what will happen if you replace the incompressible couplings by large, easily compressible rubber balls? What will then happen if you increase the loading of the trucks?

1.4.4 P-waves and S-waves

So far you will have noticed that we have only described one particular kind of disturbance, in which the bits of the medium are displaced in the direction of propagation of the disturbance, where *ball A moves towards or away from ball B* (Figure 4).

Can you think of another way of displacing ball A which might cause a different kind of disturbance to travel down the line?

Yes, of course, you could push it sideways, in a direction perpendicular to the line of springs. In Figure 6 you are looking down on top of the line of springs and balls. Ball A is shaken from side to side with the consequence that a shake-wave

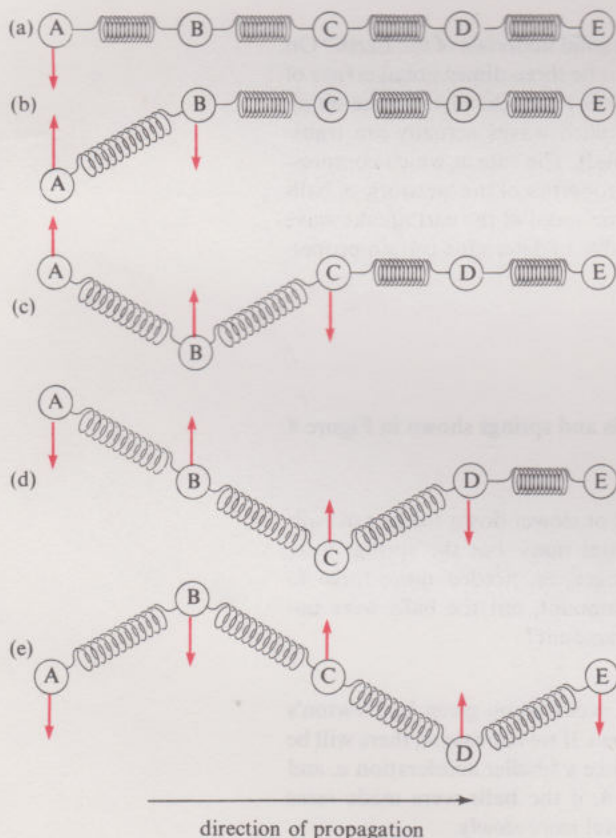


FIGURE 6 Sideways propagation of a disturbance down the line of balls and springs in Figure 4. Here you are looking down on top of the line, and ball A is moved from side to side. The stages (a-e) show the effect down the line of one complete 'cycle' of transverse movement of A. (Red arrows indicate forces.)

pulse moves down the line and the diagram shows how the wave is developed at five successive instants in time. In Figure 6e, the first sideways shake at A is about to affect ball E as a new shake pulse commences at A. We call this type of motion

transverse because the wave motion is at right angles, or *transverse*, to the direction of propagation. You could set up *transverse wave* motion for yourself by tying a clothes-line or skipping rope to a table leg, holding it fairly taut, and vibrating it from side to side. It doesn't matter, in fact, whether you shake it from side to side, up and down or any other way: the shaking motion implies that you are generating transverse waves.

In three dimensions, shake waves can be transmitted through any medium so long as it shows *any* resistance at all to being bent, that is, so long as it has any *rigidity*. So by now you should be able to visualize two kinds of wave being propagated through a solid medium, whether it is the wood of your table-top, or the solid rock of the Earth through which earthquake waves are transmitted:

1 The first type of wave we call *P-waves*, the *push-pull* variety, in which the disturbance consists of a pressure pulse, and the motion of the material of the medium takes place in the direction of propagation.

2 The other type we call *S-waves*, the *shake* or *shear* variety, where the disturbance is a wiggle pulse and the motion of the material is transverse with respect to the direction of propagation.

We call them *waves* because they appear to behave in a roughly similar way to waves we are familiar with, such as water waves. In both cases, the particles oscillate about their mean positions as the disturbance passes and they eventually come to rest exactly where they started from. So, as with water waves, the medium does not travel along during wave motion: what does travel?

1.4.5 What travels during wave motion?

The most important thing to keep in mind about these waves is that what is transmitted, what travels, through the medium is a *state of disturbance*, or *displacement*.

Is that really all that is transmitted—a mere *state* of something, an abstract notion? There is nothing abstract, you will say, about what happens when an earthquake wave hits a town! To produce the amount of damage to buildings caused by some earthquakes you would need a lot of demolition machines or a lot of high-energy explosives! Evidently, something more real, more tangible than a mere 'state' of compression or vibration is transmitted, and that something has effects that would require machines (which would use up fuel) or chemical explosives to simulate. We call that something *energy*. You will learn more about energy in Unit 8. For the moment, it is quite sufficient to think of it as something you need to move buildings about or to make corks bob up and down on water (Section 1.4.1), or to make a clothes-line shake from side to side (Section 1.4.4). Earthquake waves transmit a state of disturbance, and with it they transmit *energy*. You can think of energy being transmitted from particle to particle in the rock just like the passage of energy from ball to ball down the models we were using in preceding Sections.

Before moving on to the next Section, where you will learn how earthquake waves are generated and propagated through the Earth, just remind yourself of one of the most important conclusions from Section 1 of this Unit (cf. Section 1.4.3):

We expect waves to travel faster through a more rigid medium and slower through a more massive, hence a denser, medium.

Now that you have completed Section 1 of this Unit you should be able to:

(a) Indicate that you have some idea of the meaning of certain terms, namely: seismology, seismometer, seismic waves, a *thought experiment*, wave propagation, a compression pulse, a compression wave, a transverse wave, *P-waves*, *S-waves*, energy. (We have italicized the ones we think are most important at this stage.) (SAQ 1)

(b) Give a simple explanation for the difference between the average density of the Earth and the average density of rocks found at the Earth's surface. (ITQ 1, SAQ 2)

(c) Explain, by giving an example, that what is propagated by surface waves on water, or a seismic wave, or a compression wave down a row of balls and springs is not the medium itself but a state of disturbance and also energy. (Experiments in Sections 1.4.1 and 1.4.4, SAQ 3)

(d) Explain how the velocity of a wave disturbance will vary when the stiffness or density of the propagating medium is changed. (ITQ 2, SAQ 3)

Here are some self-assessment questions that you can use to check, or re-check, your understanding of these Objectives.

SAQ 1 Insert into the blank spaces (A–I) of the following paragraphs whichever of the terms listed below (1–9) that seem most appropriate.

Although the interior of the Earth is inaccessible to direct observation, a great deal is known from the study of (A) . These are of two different types, one compressional and one transverse. The former are called (B) and the latter (C) . In both cases, the term *wave* is used because of the similarities in the process of propagation of these seismic waves and that of the more familiar surface waves on water.

One way of visualizing the processes involved and making some qualitative or semi-quantitative predictions about (D) is to do what is called a (E) involving a simplified or idealized model. In terms of such a model one can visualize the propagation of a transverse displacement pulse in the case of S-waves and a (F) in the case of P-waves.

The study of earthquake waves, which is the science of (G) , has progressed rapidly during the past thirty years. One important reason for this is the development of very sensitive (H) to detect small underground nuclear explosions at great distances. It is of course the (I) released in such explosions that is transmitted by the seismic waves through the body of the Earth to the detecting station.

List of terms

- | | | |
|---------------------|----------------|----------------------|
| 1 compression pulse | 4 seismology | 7 S-waves |
| 2 energy | 5 seismometers | 8 thought experiment |
| 3 seismic waves | 6 P-waves | 9 wave propagation |

SAQ 2 As you will see from later Sections of this Unit, the interior of the Earth is divided up into four layers: an 'inner core'; an 'outer core'; a 'mantle'; and a 'crust'. Best available data on the average densities and volumes of these four layers are summarized in Table 1. Are these figures consistent with an average density for the Earth as a whole of $5.5 \times 10^3 \text{ kg m}^{-3}$?

TABLE 1 Densities and volumes of layers of the Earth

	Inner core	Outer core	Mantle	Crust
Volume of layer as percentage of total volume	1	15	83	1
Average density (10^3 kg m^{-3})	14.0	10.6	4.5	2.8

Hint First derive an expression for the total mass of the Earth in terms of density and volume, and then for each of its component layers as in Section 6.1 of Unit 3.

SAQ 3 (a) Use Newton's second law of motion (Unit 3, Section 3.6) to explain (1–2 sentences) why, when an earthquake enters a medium of lower density, its velocity increases, assuming other relevant properties remain the same.

(b) Earthquake waves transmit substantial amounts of energy from the source of the earthquake to all points on the Earth's surface. State (1–2 sentences) the evidence for the truth of this assertion.

2 More about earthquake waves

In Section 1 we discussed S-waves and P-waves in terms of rows of balls and springs: it is not too hard to imagine that the rock of the Earth's crust is a fairly stiff and rigid medium that should be able to carry both kinds of wave quite well. But, in this Section, we ask: what is the process that triggers the earthquake?

2.1 Earthquake mechanisms

Almost all great earthquakes are associated with movements along large cracks in the Earth's crust called *faults*. In Unit 6 we shall be showing you why faults occur where they do, but it is worth noting here that, if it were possible to observe the Earth's surface for a sufficiently long time, you would find that large segments move relative to each other. For example, the San Andreas fault near San Francisco (Figure 7) is a place where a *shearing force*, which pushes adjacent segments in opposite directions, tends to cause two segments to move laterally past each other. Although the shearing force or stress is continually applied to the blocks either side of the fault, continuous movement does not occur. What seems to happen is that the shearing force causes the rocks to be *strained*, or compressed (a mathematical definition of strain is given in Section 3.1) and, as strain is progressively accumulated, a time is eventually reached when something has to give. The rocks break and jerk past one another on either side of the fault releasing the accumulated strain partly as heat, due to friction along the fault plane, and more importantly as seismic P-waves and S-waves.

shearing force

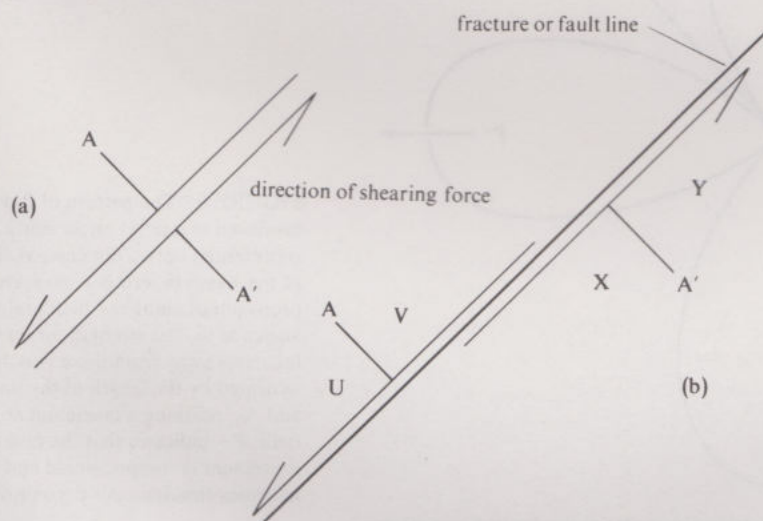


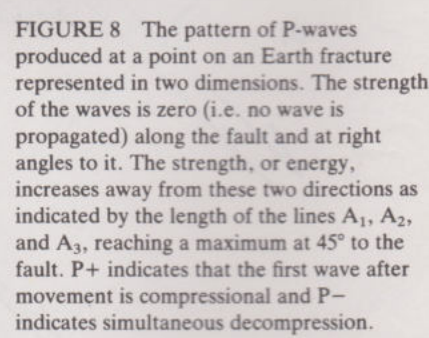
FIGURE 7 Two-dimensional representation of the San Andreas fault plane at the Earth's surface: (a) before fracture (b) after fracture. The line A-A' is drawn on the surface to illustrate relative movement. The labels U, V, X and Y refer to the question in text.

The San Andreas fault, represented in Figure 7, has accumulated strain since the last major earthquake in 1906, when a maximum horizontal displacement of 7 metres devastated San Francisco. On this occasion, the fault broke over a distance of 300 kilometres and, although smaller displacements and smaller lengths are usually involved, the amount of energy released over a short time is nevertheless extremely large.

Now you have learnt that slip on fault planes is the most important source of earthquakes, ask yourself what forces, compressional or extensional, apply to the rock medium around the fault plane when movement occurs. Consider the four places U, V, X and Y on Figure 7b.

Take the point where the line AA' crosses the fault plane as the centre of an earthquake. At the moment of fracture, the material at Y will be pushed or compressed in the direction of motion, whereas that at X will be decompressed or extended. On the other side of the fault, movement is in the opposite sense, so rock at U is compressed and at V it is decompressed. Forces are compressional at U and Y and extensional at V and X.

The result of these opposing motions for the propagation of seismic waves is as follows. P-waves are not transmitted in the direction of the fault plane because the relative motions immediately adjacent to that plane are in opposite directions and so they cancel each other out. In the direction at right angles to the fault plane, the



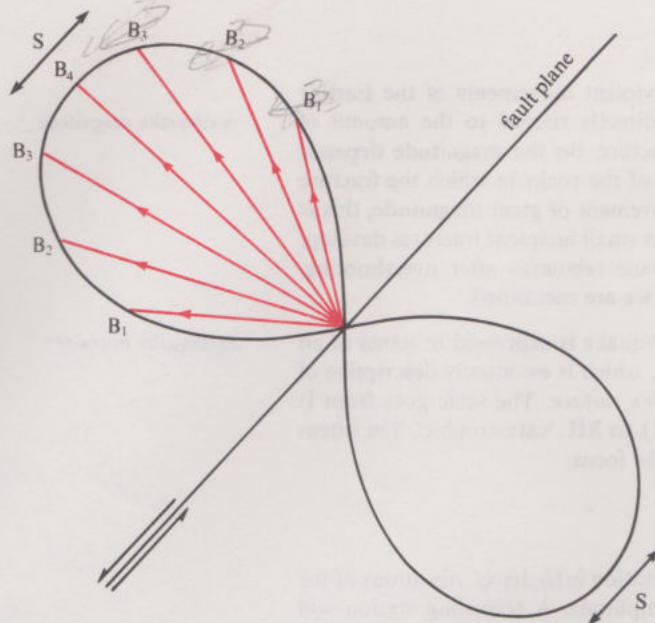


FIGURE 9 S-wave energy distributions about a point on the fault plane. The arrows labelled S indicate the direction of movement whereas those labelled B₁ to B₄ indicate the strength of the shake in these directions.

sufficiently accurate to consider all the waves from an earthquake as coming from a point source and we call this the *focus*. The focus is often tens of kilometres deep within the Earth and, in that case, we must also think of the energy lobes as extending upwards too, towards the Earth's surface. It is often useful to have a surface reference point for earthquake foci, and a line drawn from the centre of the Earth, through the focus, cuts the Earth's surface at a point known as the *epicentre*. Thus, the star in Figure 2 is the focus, and the point above it at the surface, the epicentre.

earthquake focus

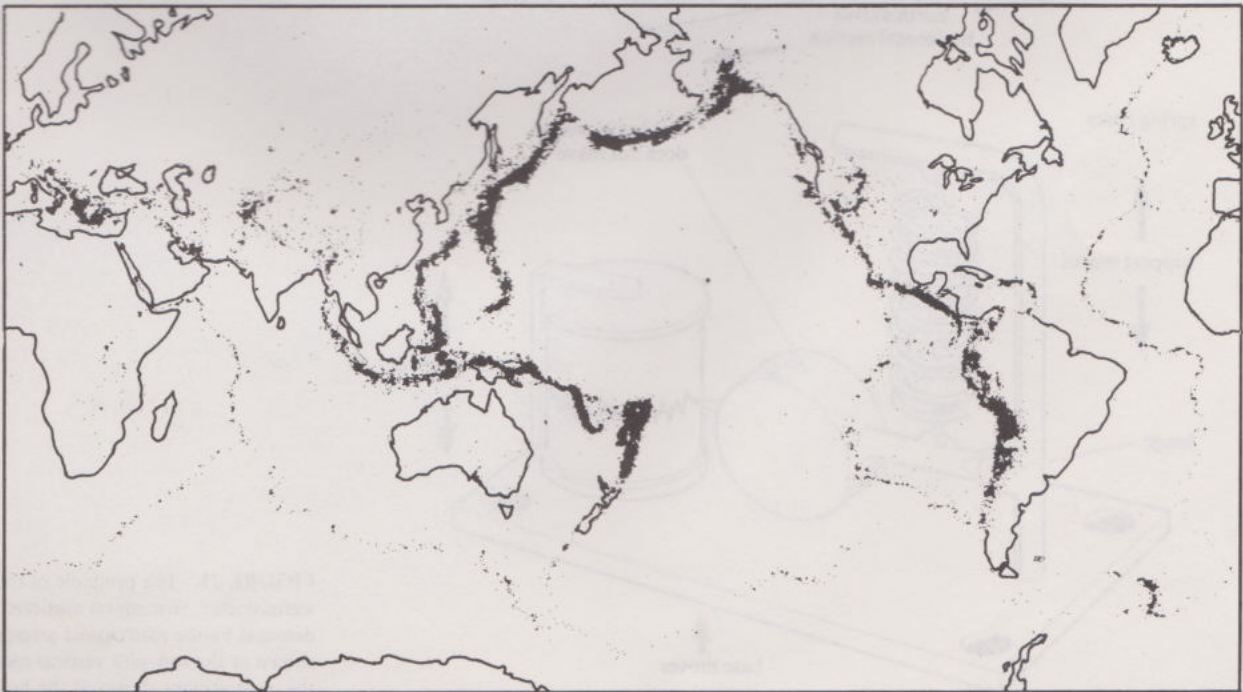
earthquake epicentre

seismic zone

FIGURE 10 Map showing the distribution of earthquake epicentres during the present century.

2.2 Earthquake zones

The majority of earthquakes originate in well-defined *earthquake or seismic zones*. In Figure 10 you can see that there are two main continental seismic zones, one bordering the Pacific, and a broader zone corresponding roughly to the Alpine-Himalayan mountain chain. These are both mountainous regions where the outer part of the Earth is being actively deformed. You can also see that seismic belts occur in the oceans; these again are associated with mountainous areas, submarine ranges whose peaks are hundreds of metres below sea-level. These earthquake zones actually mark the boundaries of the mobile segments, usually called plates, which we shall consider in Unit 6.



2.3 Earthquake magnitude and intensity

Earthquakes vary from mild quaverings to violent movements of the Earth's surface. The *magnitude* of an earthquake is directly related to the amount of energy released when over-strained rocks fracture. So the magnitude depends essentially on the inherent breaking strength of the rocks in which the fracture occurs. Although there is usually a single movement of great magnitude, this is often preceded by low-magnitude *foreshocks*, as small incipient fractures develop, and succeeded by *aftershocks* if the fault plane rebounds after overshooting. However, it is the *principal shock* with which we are concerned.

The *intensity* of the vibrations due to an earthquake is expressed in terms of an arbitrary, 12-point, scale devised by Mercalli, which is essentially descriptive of the *damage done* at some point on the *Earth's surface*. The scale goes from I: 'instrumental' (only detectable by instruments), to XII: 'catastrophic'. The intensity decreases with increasing distance from the focus.

earthquake magnitude

earthquake intensity

2.4 The recording of earthquake waves

The task of the seismometer at the recording station is to detect vibrations of the Earth that may be as small as 10^{-6} m in amplitude. A recording station will usually have at least three seismometers—one to pick up vibrations in the verti-

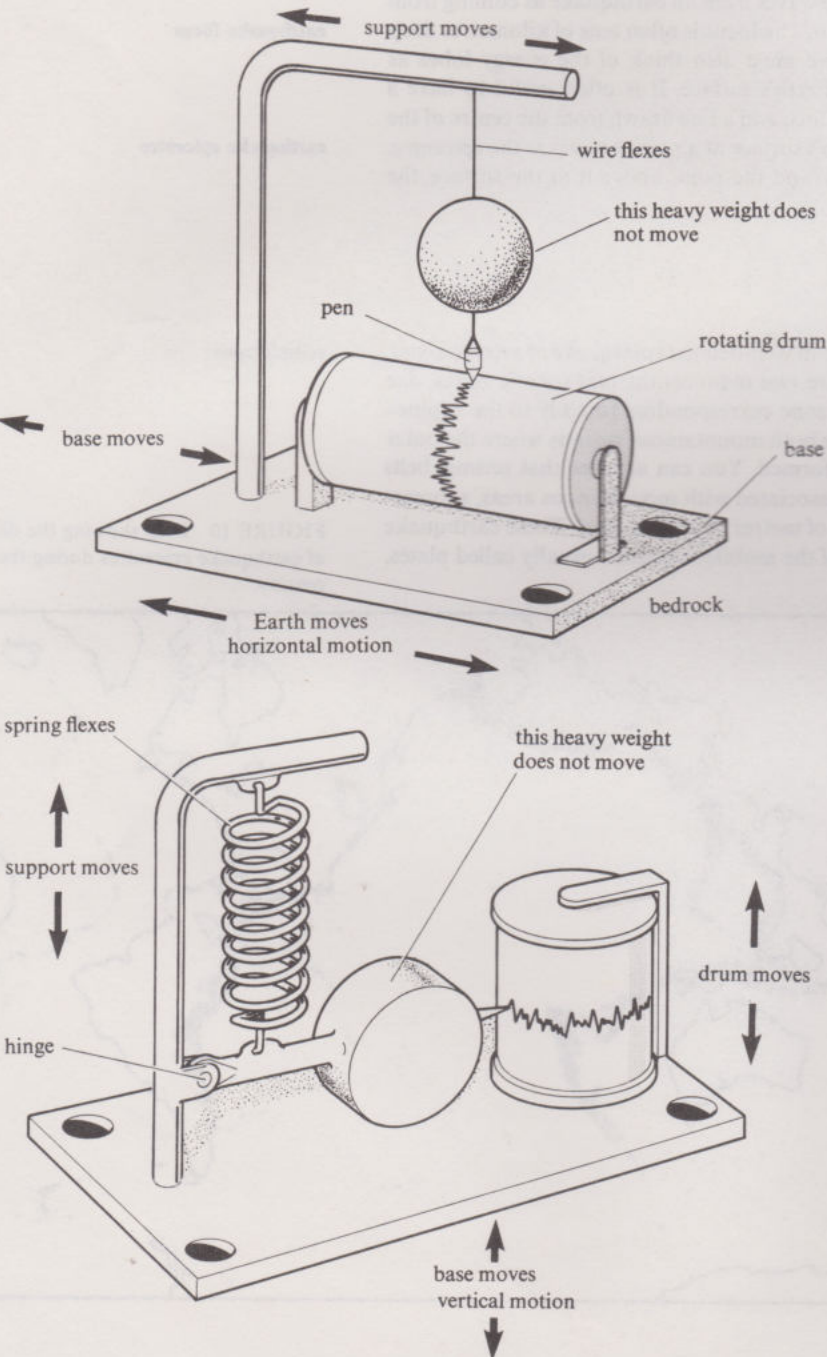


FIGURE 11 The principle of the seismometer. Horizontal motions are detected by the mechanical arrangement shown at the top, and vertical motions by the arrangement shown at the bottom.

cal direction and two others to record in east-west and north-south directions, all three being known as a seismic set. The principle on which the seismometer works is that part of it, a heavy weight, remains stationary while the waves pass, whereas the rest of the apparatus, in direct contact with the ground, vibrates. The magnitude and frequency of the vibrations are then recorded as a wiggly trace on a time-correlating device; this record is a *seismogram*. Modern seismometers are sophisticated devices that detect, amplify, filter and record the motions of the Earth, but the principle is exactly the same as that illustrated in Figure 11. Some of these points are illustrated in the TV programme associated with this Unit (TV 04)

Now that you have completed Section 2 of this Unit you should be able to:

- Show that you understand the meaning of: a shearing force, an earthquake focus, an earthquake epicentre, a seismic zone, the magnitude and intensity of earthquakes, a seismogram. (SAQ 4)
- Explain in simple terms why a shearing fracture of the rock of the Earth's crust can set up both P-waves and S-waves. (SAQ 5)
- Explain in simple terms the variation in strength of P-waves and S-waves with respect to the direction of fault movement. (SAQ 5)
- Distinguish, on a map of the Earth, an approximately correct distribution of seismic zones from a grossly incorrect one. (SAQ 6)
- Describe the basic principle of a simple seismometer. (SAQ 7)

The following SAQs will test whether you have achieved these Objectives.

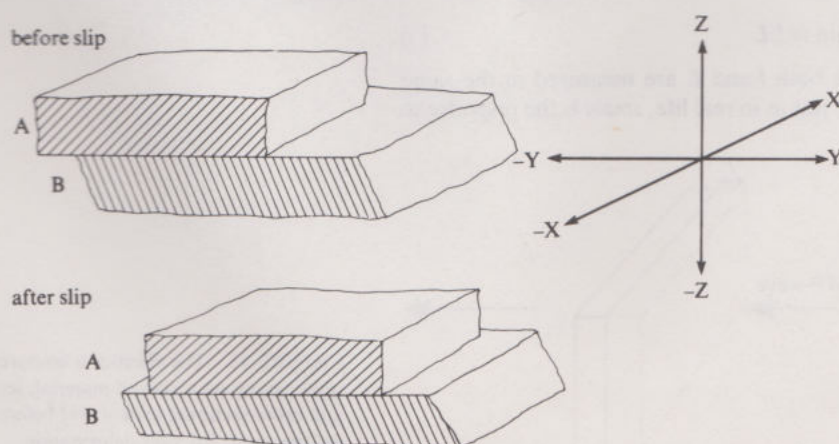
SAQ 4 (i) Which of the forces shown in Figure 12, deforming a cube of material, is a shearing force?

(ii) Which of the following statements are correct?

- The focus of an earthquake is the point from which the earthquake wave originates.
- A seismic zone is a region on the Earth's surface where earthquakes occur most frequently.
- The intensity of an earthquake is a direct measure of the energy released at its focus.
- The epicentre of an earthquake is the point on the Earth's surface vertically above the point where the earthquake occurs.
- The intensity of an earthquake is a quantitative measure of the damage it causes of a particular point on the Earth's surface.
- A seismogram is an instrument for detecting earthquake waves.

SAQ 5 Two slabs of rock A and B suddenly slip sideways relative to each other in the Y direction and along the X-Y plane as illustrated in Figure 13. Both slabs are entirely surrounded by other rock and A is vertically above B.

- In which of the directions X, -X, Y, -Y, Z, -Z will there be propagation, at some distance from the fault, of (i) P-waves, (ii) S-waves?
- In which directions relative to the frame of reference XYZ will (i) the P-waves, (ii) the S-waves have maximum strength?



seismogram

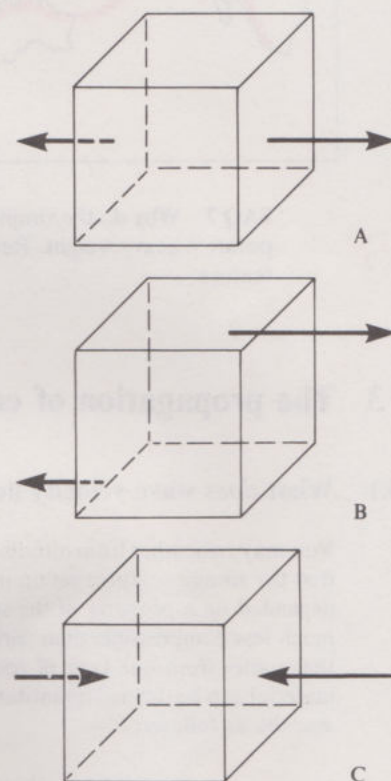


FIGURE 12 For use with SAQ 4.

FIGURE 13 For use with SAQ 5.

SAQ 6 Figure 14 shows possible distributions of earthquake epicentres. Which one of the five seismic zones shown here is correctly located on the map?

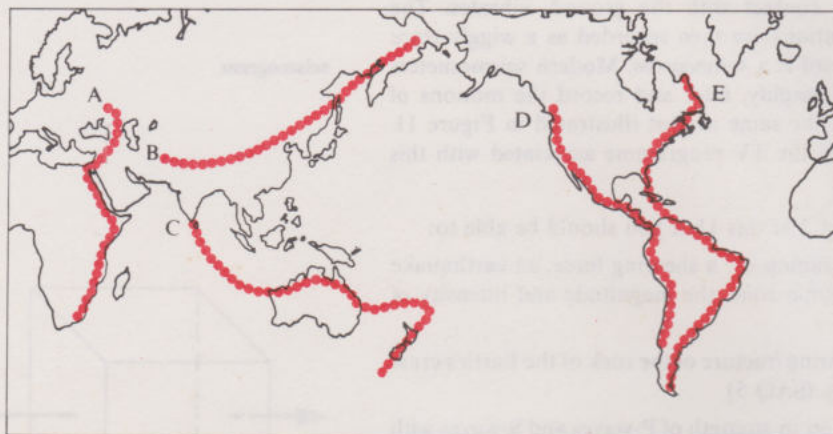


FIGURE 14 Hypothetical seismic zones for use with SAQ 6.

SAQ 7 Why do the simple seismometers shown in Figure 11, p. 16 incorporate a heavy weight. Recall from Unit 3 the basic law that underlies this feature.

3 The propagation of earthquake waves

3.1 What does wave velocity depend on?

You may remember from our discussion of the ball-and-spring model of materials that the amount of force set up in each spring by a given displacement of the ball depended on a property of the spring that we called its *stiffness*. Now, rocks are much less compressible than springs, but they still have a characteristic 'stiffness' that varies from one type of rock to another. This characteristic property of a material can be defined quantitatively in terms of a modulus of elasticity or *elastic modulus* as follows:

$$\text{elastic modulus} = \frac{\text{stress}}{\text{strain}} \quad (1)$$

The term *stress* means *force per unit area*. In the case of a P-wave travelling through a solid, the forces are *compressional*, so we have *compressional stress*. Imagine a slice cut at right angles to the direction of wave propagation (Figure 15a). If its cross-sectional area is A and if the compressional force acting on the slice is F , then:

$$\text{stress} = F/A \text{ N m}^{-2} \quad (2)$$

If the original length of the slice is L in the direction of propagation, and the pulse compresses the slice by an amount l (Figure 15b), then the degree of deformation is l/L , which is called the *strain*:

$$\text{strain} = l/L \quad (3)$$

This is a dimensionless ratio because both l and L are measured in the same dimensions—length. Remember that, just as in real life, *strain* is the response to *stress*!

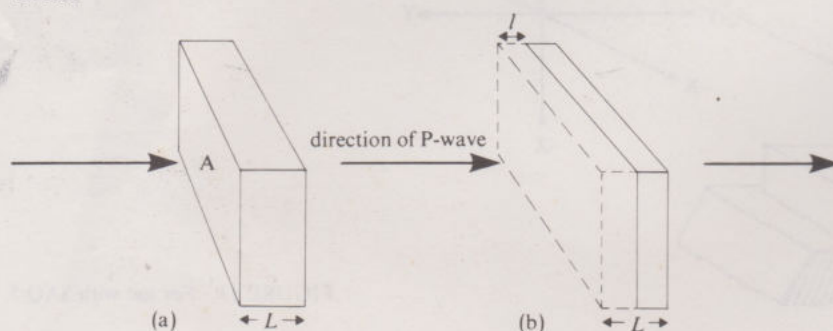


FIGURE 15 The effect of a compressional pulse acting on a slab of material, length L and cross-sectional area A : (a) before deformation; (b) after deformation.

Substitution into equation 1 gives an expression for the elastic modulus, which in this case is the *incompressibility** of the medium, which we have considered in a *longitudinal* sense (i.e. in the direction of propagation):

$$\text{incompressibility} = \frac{\text{longitudinal stress}}{\text{longitudinal strain}} = \frac{F/A}{l/L} \text{ N m}^{-2} \quad (4)$$

The treatment in Figure 15 is not strictly applicable to P-waves passing through rock because, unlike rock, the slab in Figure 15 can expand sideways when it is stressed. The specific elastic modulus that applies more literally to rock under longitudinal stress, where lateral expansion is prevented, is called the *axial modulus* and is represented by the Greek letter ψ (pronounced 'psi'):

$$\text{axial modulus } \psi = \frac{F/A}{l/L} \text{ N m}^{-2} \quad (5)$$

Now we are ready to use the results of our thought experiments about the dependence of pulse-propagation velocity in ball-and-spring models on spring strength (now related to ψ) and mass (related to density).

The following proof of the relationship between seismic P-wave velocity, axial modulus and density is non-essential and we suggest you take the result (equation 16) for granted on a first reading, especially if you are short of time. The proof is included for interest in case you wish to follow it.

(Throughout S101, whenever there is a passage of non-essential reading such as this, we shall indicate its start with a single star ★ in the margin, and its end with a double star ★★. So, if you are short of time, continue reading from the double star.)

Recall from Section 1.4.3 the relationship between propagation velocity, spring stiffness and ball mass (m). Our first conclusion was that stiffer springs cause an increase in propagation velocity (v_p). Now, stiffer springs require more force to produce a given deformation and so represent greater values of axial modulus (see equation 5). The other conclusion was that less massive balls cause velocities to increase. ★

These conclusions can be expressed as follows:

$$v_p \text{ increases IF } \psi \text{ increases} \quad (6)$$

$$v_p \text{ increases IF } m \text{ decreases} \quad (7)$$

We can extend equation 7 by remembering that there is a characteristic property of a material that tells you how much mass there is in a given volume of it. Can you remember what this is?

It is, of course, the *density*, which is usually represented by the Greek letter ρ (pronounced 'rho') and measured in kilograms per cubic metre. So we reinterpret equation 7 as follows:

$$v_p \text{ increases IF } \rho \text{ decreases} \quad (8)$$

which is the same as saying:

$$v_p \text{ increases IF } \left(\frac{1}{\rho}\right) \text{ increases} \quad (9)$$

Combining conclusions 6 and 9 together:

$$v_p \text{ increases IF } \psi \text{ increases OR IF } \left(\frac{1}{\rho}\right) \text{ increases}$$

which can be stated more succinctly as:

$$v_p \text{ increases IF } (\psi/\rho) \text{ increases} \quad (10)$$

* Note that we use the term 'incompressibility' here rather than 'compressibility' because l appears at the bottom of the equation. Suppose we fix F , A and L , and then compare l for a lump of rock and a lump of cheese. For the rock, l will be small, hence a large incompressibility, i.e. the rock is not easily compressed, but for the cheese l will be large because we have a small resistance to compression or small incompressibility.

Now the conclusion stated by equation 10—that v_p increases if the ratio (ψ/ρ) increases—is based on the thought experiment that established the essentially qualitative relationships stated in equations 6 and 9. But we cannot tell from this thought experiment whether the velocity should be simply proportional to (ψ/ρ) , or to the square of (ψ/ρ) , or to its cube, or whether there is some other, more complicated, way in which v_p increases if (ψ/ρ) increases.

Let us start by assuming that the relationship between v_p and ψ/ρ is the *simplest possible* one that would give the required semi-quantitative result, that v_p increases if ψ/ρ increases, namely that:

$$v_p \propto (\psi/\rho) \quad (11)$$

where $\propto$ means 'is proportional to'.

It is always best, in constructing a model of whatever kind, to *start* with the *simplest possible* one. Now we can try to check whether equation 11 is the correct relationship.

We do this by analysing the dimensions (Unit 2, Section 2.5.1) used to measure the three quantities:

v_p is a velocity, units: m s^{-1} ; dimensions: $[\text{length}]/[\text{time}]$

ψ is a modulus, units: N m^{-2} ; dimensions: $[\text{mass}]/([\text{length}] \times [\text{time}^2])$

ρ is a density, units: kg m^{-3} ; dimensions: $[\text{mass}]/[\text{length}^3]$

Just to explain how we arrive at the dimensions of ψ , remember that the unit of force, 1 newton = $1 \text{ kg} \times 1 \text{ m s}^{-2} = 1 \text{ kg m s}^{-2}$. So, if we replace mass by M, length by L and time by T (see HED, Section 4.3), the ratio ψ/ρ would be measured in the dimensions:

$$\left[\frac{\psi}{\rho} \right] = \frac{[\text{ML}^{-1}\text{T}^{-2}]}{[\text{ML}^{-3}]} = [\text{L}^2\text{T}^{-2}] \quad (12)$$

But we know that the left-hand side of equation 11 is a velocity, with dimensions $[\text{LT}^{-1}]$. Evidently, the simplest possible assumption was a bit too simple because this leads to $[v_p] = [\text{L}^2\text{T}^{-2}]$.

If the assumption that $v_p \propto \psi/\rho$ leads to the wrong conclusion that $[v_p] = [\text{L}^2\text{T}^{-2}] = [(\text{LT}^{-1})^2]$, what relationship between v_p and (ψ/ρ) will lead to the right conclusion that $[v_p] = [\text{LT}^{-1}]$?

The answer is: $v_p \propto (\psi/\rho)^{1/2}$ (13)

for then, $[v_p] = \left[\left(\frac{\psi}{\rho} \right)^{1/2} \right] = [(\text{L}^2\text{T}^{-2})^{1/2}] = [\text{LT}^{-1}]$

Equation 13 is the same as:

$$v_p \propto \sqrt{\frac{\psi}{\rho}} \quad (14)$$

and this may be written as:

$$v_p = k \sqrt{\frac{\psi}{\rho}} \quad (15)$$

where k stands for a constant of proportionality. Now, if we measure v_p in m s^{-1} and (ψ/ρ) in $(\text{m s}^{-1})^2$, the constant of proportionality in equation 15 will be dimensionless** and equal to unity, i.e. $k = 1$, so:

$$v_p = \sqrt{\frac{\psi}{\rho}} = \sqrt{\frac{\text{axial modulus}}{\text{density}}} \quad (16)$$

which says that the velocity of P-waves (v_p) is equal to the square root of the axial modulus (ψ) divided by the density (ρ). This is an extremely important relationship which will be used throughout the rest of this Unit: it tells us that if we can

* If you are uncertain about the terms 'is proportional to' and 'constant of proportionality', which appears below, refer to MAFS, Block 2.

** See HED, Section 7.2.



MAFS 2

MAFS 2

HED 4.2 ★★

measure the P-wave velocity for part of the Earth's interior, then we can begin to say something about the relative compressibilities and densities of the materials there.

So far we have considered only P-waves. Now what about S-waves. The reasoning leading up to equation 16 can be repeated for S-waves, except that, instead of the axial modulus ψ , which has to do with *compressional* stress, we should use a different elastic modulus which has to do with the stress accompanying *transverse* movement transmitted by S-waves. This kind of stress is called *shearing stress* and the lateral deformation it produces is called a *shearing strain*. The ratio of stress to strain in this case is called the shear modulus or *rigidity modulus*.

shearing stress
shearing strain
rigidity modulus

$$\text{rigidity modulus } (\mu) = \frac{\text{shearing stress}}{\text{shearing strain}} \quad (17)$$

(the Greek letter μ is pronounced 'mew')

Remembering equation 16 for P-waves, what relationship would you expect to find between rigidity modulus, density and S-wave velocity?

If S-waves behave like P-waves, we might get away with substituting the rigidity modulus for the axial modulus. In fact, this is the relationship which is found experimentally:

$$v_s = \sqrt{\frac{\mu}{\rho}} \quad (18)$$

which is an equally important relationship in dealing with the Earth's interior. You should try to remember equations 16 and 18, which we shall also develop practically in TV 04.

Rocks, like most solids, resist compression better than they resist shear, that is, they need more compressional stress (more force) to produce a certain compressional deformation (longitudinal strain) than they need shearing stress to produce a comparable lateral deformation (shearing strain). Thus, for an average rock, the axial modulus is about $3 \times 10^{11} \text{ N m}^{-2}$ and the rigidity modulus is about $8 \times 10^{10} \text{ N m}^{-2}$.

ITQ 3 Given these values for elastic moduli and the Earth's density ($\rho = 5.5 \times 10^3 \text{ kg m}^{-3}$), determine approximate values for the velocity of P-waves and S-waves in rocks.

P-waves travel faster than S-waves because the resistance of a rock to compression is greater than its resistance to shear. Going back to balls and springs: it is easier to bend 'springs' side-ways than to compress them!

ITQ 4 Bearing in mind that sound waves obey the same sort of relationship as in equation 16, explain in your own words why sound travels progressively faster in air, wood, water and a metal pipe.

Now another question: can both P-waves and S-waves be transmitted through a *liquid*? If you cannot answer this by simple reasoning about the terms ψ and μ , then go back to the sink full of water. Take a small plate and hold it in the water at one end and, at the other, place your hand in the water with the palm facing the plate. Now create a P-wave by pushing the plate with a short, sharp jerk towards your other hand—you should feel the P-wave quite distinctly. Now, with a similar jerk, move the plate sideways creating a shear wave. Do you feel anything? So long as you have induced a true shearing motion then you should not feel any motion. *Liquids have no appreciable rigidity* and do not resist shearing deformation: they cannot be bent. You might argue, and reasonably so, that a viscous liquid—like thick custard—has rigidity, but we challenge you to observe shear deformation across a vat of custard several kilometres wide!

So a substance with a rigidity modulus close to zero cannot be bent because it will not resist shear. Thus, S-waves will not pass through liquids, or gases for that matter. Both are compressible, have axial moduli that are not zero and, therefore, transmit P-waves. As you will see, this is an important clue to the behaviour of seismic waves when they pass through the Earth's interior.

Before moving onto Section 3.2 you should check that you can achieve the following Objectives:

- Indicate that you know the meaning of the following terms: elastic modulus, stress, compressional stress, compressional strain, axial modulus, shearing stress, shearing strain, rigidity modulus. (SAQ 8)
- Calculate the velocities of P-waves and S-waves through a solid medium given the density and the relevant elastic moduli. (ITQs 3 and 4)
- Outline the main steps in a semi-quantitative derivation of the formulae for the velocities of P-waves and S-waves through a solid medium. (This Objective relates to the non-essential section.) (SAQ 9)
- Explain in simple terms why an S-wave cannot be propagated through a liquid. (SAQ 10)

SAQ 8 Which of the following statements are true?

- The elastic modulus measures the stress produced by a given force.
- The stress on a body is the force per unit cross-sectional area of the body.
- If a body of cross-sectional area A is acted on by a compressional force F , the compressional stress is A/F .
- If, under the action of a compressional force, the length of a body is reduced by 10 per cent, this corresponds to a compressional strain of 0.1.
- When a body is deformed by forces which tend to displace successive layers of the material laterally relative to each other, the stress is called a shearing stress, and the ratio of this stress to the corresponding strain is called the rigidity modulus.

SAQ 9 Chart 1 shows some of the important steps we followed in deriving, semi-quantitatively, the formula for P-wave and S-wave velocities. Fill in the blanks A–J.

SAQ 10 Suppose an explosive charge is set off in the polar ice just above the water level, at a point such as P in Figure 16. Some distance away two seismometers A and B are placed, one on the surface of the ice and the other in the water just below the ice.

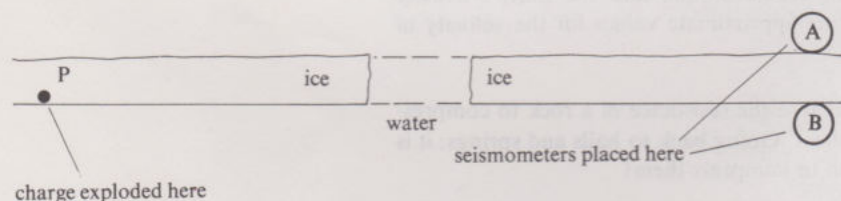
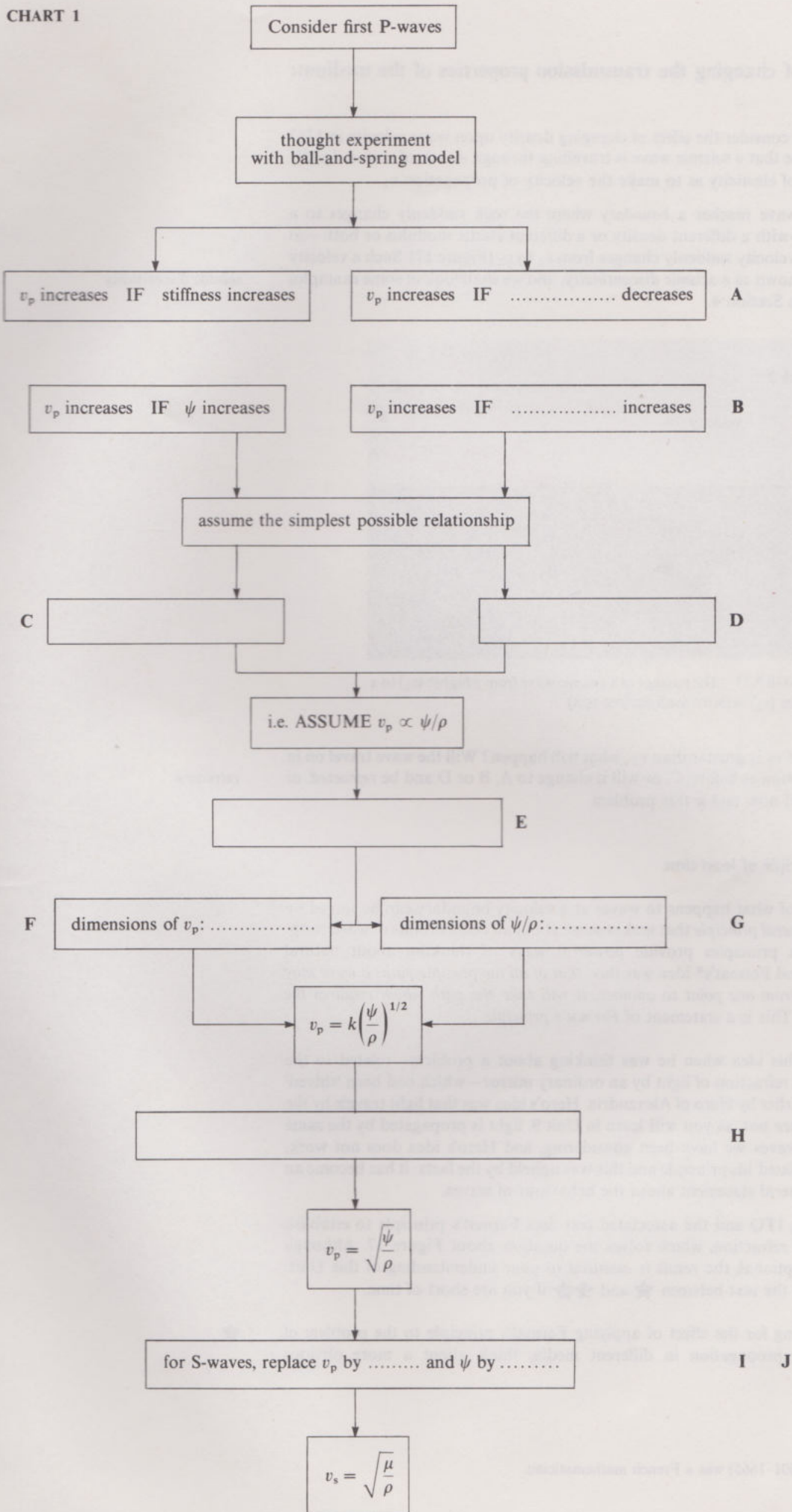


FIGURE 16 For use with SAQ 10.

- Would you expect both seismometers to register both P-waves and S-waves.
- Consider the relative resistance to compressional deformation of ice and water. Assuming that the densities of ice and water are identical, decide which seismometer would be the first to record P-waves.



3.2 The effect of changing the transmission properties of the medium: refraction

You started to consider the effect of changing density upon wave velocity in ITQ 4. Now suppose that a seismic wave is travelling through a rock of such a density and modulus of elasticity as to make the velocity of propagation v_1 .

Suppose the wave reaches a *boundary* where the rock suddenly changes to a different one—with a different density or a different elastic modulus or both—so that the wave velocity suddenly changes from v_1 to v_2 (Figure 17). Such a velocity boundary is known as a *seismic discontinuity*, and we shall look at some examples in the Earth in Section 4.

seismic discontinuity

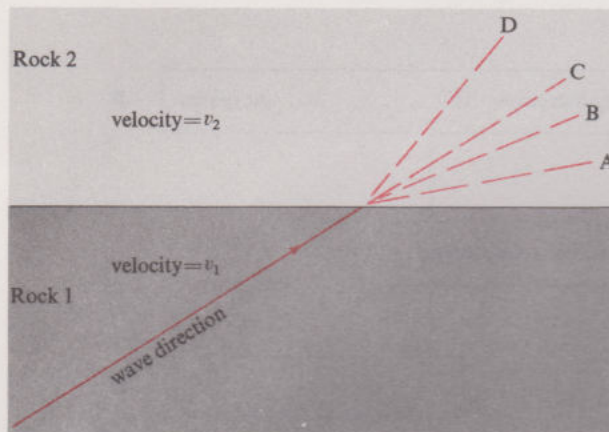


FIGURE 17 The passage of a seismic wave from a higher (v_1) to a lower (v_2) velocity medium (see text).

In Figure 17, if v_1 is greater than v_2 , what will happen? Will the wave travel on in the same direction as before, C, or will it change to A, B or D and be *refracted*, or bent? We shall now tackle this problem.

refraction

3.2.1 Fermat's principle of least time

The problem of what happens to waves at a velocity boundary can be solved by applying a *general principle* that makes sense about all observations of wave propagation. Such principles provide powerful ways of thinking about natural phenomena and Fermat's* idea was this: *that of all the possible paths a wave may follow to get from one point to another, it will take the path which requires the shortest time.* This is a statement of *Fermat's principle*.

Fermat had this idea when he was thinking about a problem—related to the reflection and refraction of light by an ordinary mirror—which had been 'solved' a long time earlier by Hero of Alexandria. Hero's idea was that light travels by the *shortest distance* but, as you will learn in Unit 9, light is propagated by the same rules as the waves we have been considering, and Hero's idea does not work. Fermat postulated his principle and this was upheld by the facts: it has become an important general statement about the behaviour of waves.

The following ITQ and the associated text uses Fermat's principle to establish Snell's law of refraction, which solves the question about Figure 17. Although the proof is optional, the *result is essential* to your understanding of this Unit: you may skip the text between ★ and ★★ if you are short of time.

To get a feeling for the effect of applying Fermat's principle to the problem of seismic wave propagation in different media, think about a more obvious example.

★

* P. Fermat (1601–1665) was a French mathematician.

ITQ 5 You are at the point Y (Figure 18) when you see someone drowning in the sea at point X, which, for simplicity's sake, is as far away from the water's edge as you are. There is deep water immediately across the edge—you have to start swimming as soon as you enter the water. Suppose you are an excellent swimmer and can travel half as fast through the water as you can over the ground. If you would rather choose some other ratio, like a third or a quarter, it will make no difference. We suggest half only because two is an easy number to multiply by.

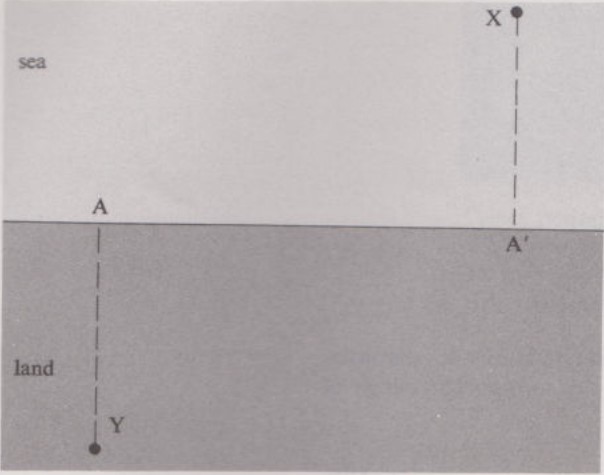


FIGURE 18 Apply Fermat's principle to decide what is the shortest time to get from Y to X. The distances YA and XA' are 50 metres and AA' is 100 metres.

Now the problem is: what path should you follow to get from Y to X in the shortest possible time? Sparing a thought for the poor chap drowning in the water, we suggest you adopt an 'experimental' approach to this problem, making a scale drawing like Figure 18 on a sheet of graph paper, measuring the lengths of various possible paths with a ruler, calculating the times assuming, for instance, a running speed of 4 m per second and hence a swimming speed of 2 m per second, and finding by trial and error, or by drawing a graph, the path that takes the least time. When you have found that path, measure the angles it makes with the perpendicular to the water line, i.e. the angles θ_1 and θ_2 in Figure 19.

The dotted line in Figure 19 is drawn at right angles to the interface—we call it a *normal*—and it is the angle between the path directions and the normal that you should measure. Incidentally, does the path we have drawn in this Figure seem a sensible one? We tried this out for ourselves, and you will find the results in the ITQ answer.

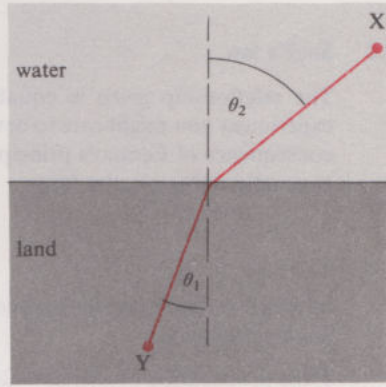


FIGURE 19 The angles θ_1 and θ_2 are the ones you should measure.

In the light of the ITQ you have just completed, recall that $v_1/v_2 = 2$ and that $\theta_1 = 56^\circ$ and $\theta_2 = 24.5^\circ$, as shown in Figure 20. Unfortunately, there does not seem to be any obvious relationship between these angles and velocities; certainly 56° is not $2 \times 24.5^\circ$, for instance. But now use your calculator to find the value of $\sin 56^\circ$ and $\sin 24.5^\circ$:

$$\begin{aligned}\sin 56^\circ &= 0.8290 \\ \sin 24.5^\circ &= 0.4147\end{aligned}$$

So it seems that $\frac{\sin \theta_1}{\sin \theta_2} = 2$

Can it be a coincidence that $\frac{v_1}{v_2} = 2$ as well?

You could satisfy yourself on this point by choosing another ratio for v_1/v_2 , say 3, and repeating the experiment. You would find that the values of $\sin \theta_1/\sin \theta_2$ were the same as the ratio you chose for v_1/v_2 : this holds good for any type of wave motion, including seismic waves.

To find the value of $\sin 56^\circ$, key: 56, sin. You should get the answer 0.82903757.

If your answer differs from this, you may have been working with your calculator in the radian or grad mode (indicated respectively by one or two vertical lines on the left of the display). Re-engage the degree mode by switching off and on again.



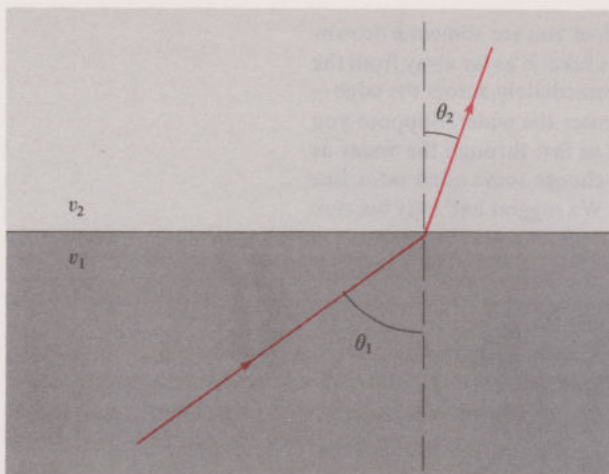


FIGURE 20 For any kind of wave motion, the angles θ_1 and θ_2 are measured between the direction of wave propagation and a normal (perpendicular) to the interface (dashed). Notice that the angle of incidence i is θ_1 and the angle of refraction r is θ_2 .

The conclusion drawn from Fermat's principle of least time is that, for the situation shown in Figure 20, where a wave passes from one medium to another of contrasting velocity:

$$\frac{\sin \theta_1}{\sin \theta_2} = \frac{v_1}{v_2} \quad (19)$$

3.2.2 Snell's law

The relationship given in equation 19 will stand up to any test of theory or experiment you might care to devise. It seems that it is a general rule—a necessary consequence of Fermat's principle. This relationship is known as *Snell's law*, and is usually written in the form:

$$\frac{\sin i}{\sin r} = \frac{v_1}{v_2} \quad (20)$$

in which we have simply replaced θ_1 by i , the *angle of incidence*, and θ_2 by r , the *angle of refraction*.

angle of incidence
angle of refraction

If the propagation of seismic waves across a boundary where the velocity changes, as in Figure 17, obeys Fermat's principle, then the wave directions can be related by Snell's law.

This *bending* of the wave direction is called *refraction*.

Now refer back to Figure 17 and decide which of the paths A–D is likely to apply for a seismic wave passing from rock 1 to rock 2, where $v_1 > v_2$. (> means 'is greater than', see *MAFS*, Block 4).

MAFS 4

Since $v_1/v_2 > 1$, and therefore $\sin i/\sin r > 1$, path D is the only possible solution, the angle of incidence being greater than the angle of refraction. Remember that (for angles between 0° and 90°) the sine of an angle increases as the angle increases, so if $\sin i > \sin r$, then $i > r$.

So a wave always bends towards the normal when entering a medium of lower velocity.

You should now be able to achieve the following Objectives:

- Understand the following terms: angle of incidence, angle of refraction, refraction. (SAQ 11)
- State Fermat's principle of least time. (Section 3.2.1)
- Describe a graphical method for deriving Snell's law of refraction and apply it to a wave passing through two media with different wave velocities. (Note that this is an optional Objective.) (ITQ 5)
- State Snell's law of refraction and apply it to a wave passing through two media with different wave velocities. (SAQ 12)

SAQ 11 (i) Which of the two diagrams in Figure 21 illustrates refraction?

(ii) On the appropriate diagram, label the angles of incidence and refraction.

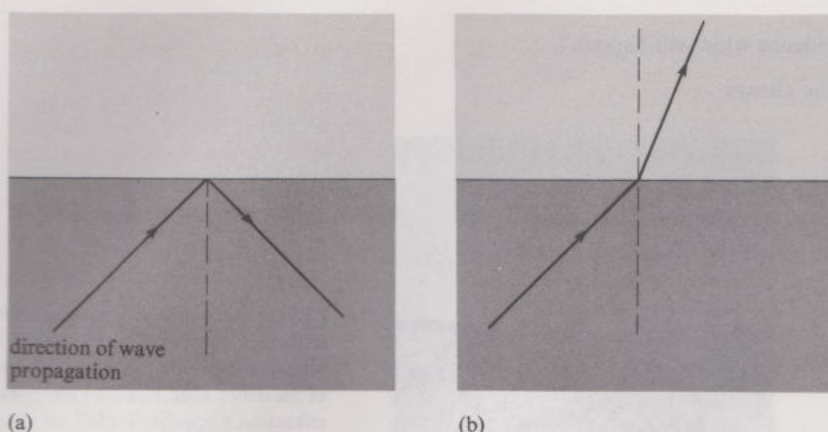


FIGURE 21 For use with SAQ 11.

SAQ 12 A wave travels with velocity v_A in a medium A and with velocity v_B in a medium B. If it travels from A to B and the angle of incidence is i , will the angle of refraction r be greater than the angle of incidence, if

(a) $v_A < v_B$?

(b) $v_A > v_B$?

(c) $v_A = v_B$?

3.3 Other transmission effects: reflection

3.3.1 The law of reflection

When a wave meets an interface between one medium and another it is not only *refracted* but also *reflected*, as illustrated in Figure 22.

Bearing in mind Fermat's principle, and that a wave stays in the same medium after reflection with *unchanged* velocity, does the reflected wave direction shown in Figure 22 seem sensible?

The *least time* taken to get between two points on the same side of an interface by reflection will be if the angles made by the incident and reflected waves to the normal are the same; so the direction shown is not sensible. In this case, where there is only one velocity involved, the least-time path is also the least-distance path (see also Figure 24).

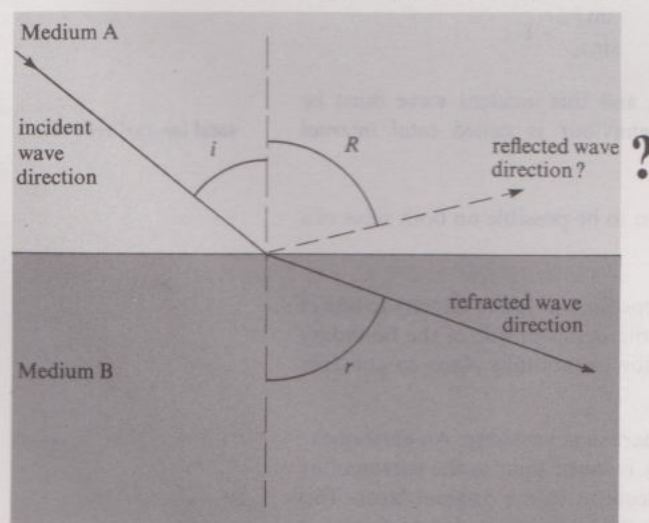


FIGURE 22 Refraction and reflection from an interface.

Thus angle of incidence = angle of reflection
 $i = R$

(21)

This statement is sometimes called the *law of reflection*.

3.3.2 Total reflection

Suppose (as in Figure 23) you have a wave travelling from a medium A towards an interface with a medium B and that the velocity in A is *lower* than the velocity in B.

As you increase the angle of incidence what will happen?

Figure 23 and its caption give the answer.

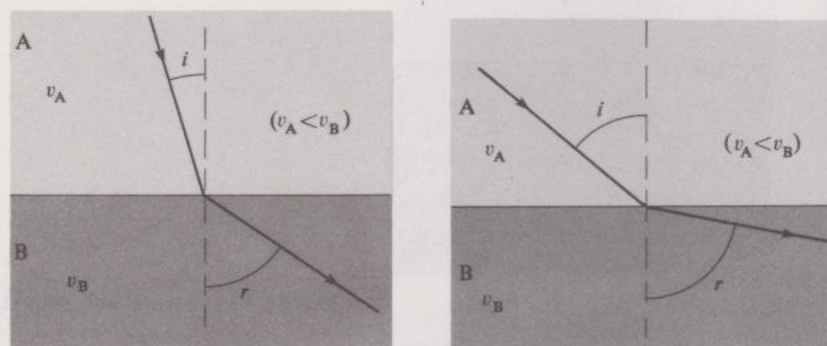


FIGURE 23 The first diagram shows refraction as a wave passes from a low to a higher wave velocity medium. As the angle of incidence i increases, so the angle of refraction r approaches 90° , as shown in the second diagram.

In Figure 23, when $r = 90^\circ$ exactly, the corresponding value of i is called the *critical angle* i_c ,

$$\text{then} \quad \frac{\sin i_c}{\sin r} = \frac{\sin i_c}{\sin 90^\circ} = \frac{\sin i_c}{1} = \frac{v_A}{v_B}$$

$$\text{i.e.} \quad \sin i_c = \frac{v_A}{v_B}$$

There is, of course, nothing to stop a wave being incident on the boundary at an angle exceeding the critical angle.

What do you expect would happen to such a wave?

When this happens the wave cannot refract into the second medium—it can only reflect back into the original medium. To emphasize this point, note that if the angle of incidence i exceeds i_c , the sine of the angle refraction would have to be greater than unity, which is not possible (see *MAFS*, Block 4):

$$\frac{\sin i}{\sin r} = \frac{v_A}{v_B} = \sin i_c$$

$$\text{hence} \quad \sin r = \frac{\sin i}{\sin i_c}$$

$$\text{so if} \quad i > i_c, \quad \sin i > \sin i_c, \quad \frac{\sin i}{\sin i_c} > 1$$

Since this is also impossible, so is refraction, and this incident wave must be entirely reflected at the boundary. Such behaviour is called *total internal reflection*.

Would you expect total internal reflection to be possible on both sides of a boundary?

No, it only affects waves trying to pass from a medium of lower velocity to one of higher velocity. If the wave approaches from the opposite side of the boundary there is no angle of incidence for which refraction cannot take place, so one does not get total internal reflection.

Total internal reflection plays an important part in seismology. An earthquake wave originating in a layer where the velocity is lower than in the surrounding layers can find itself trapped by repeated reflections in the original layer. This effect is shown in Figure 24, where you can see that the only evidence of such an earthquake, which seismometers on the surface can record, is from waves having small angles of incidence with adjacent layers.

One further point about reflection. Note that when a wave strikes an interface at less than the critical angle of incidence, the energy of that wave is divided between

critical angle

MAFS 4

total internal reflection

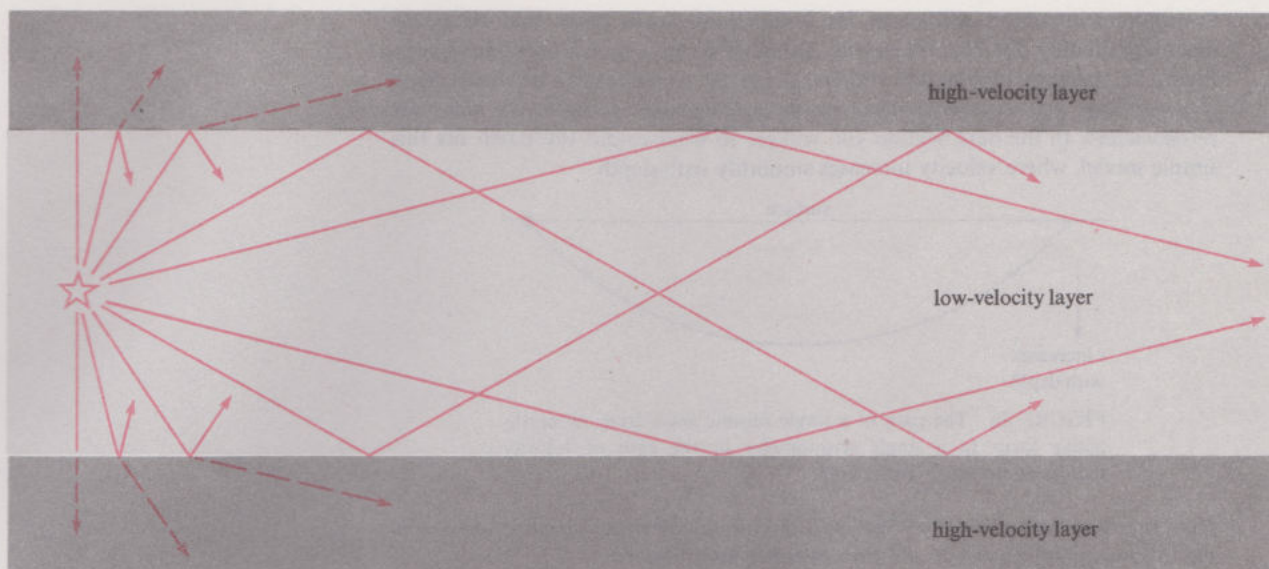


FIGURE 24 The trapping of waves in a low-velocity layer.

refraction and reflection. For a wave perpendicular to the interface ($i = 0^\circ$), the ratio of refracted (or transmitted) energy to reflected energy is a *maximum*. The strength of the reflected wave grows as the angle of incidence becomes larger until, beyond the critical angle, there is no refracted energy.

3.3.3 What happens if there is a more or less continuous change in the wave velocity?

It would be surprising if the densities and the elastic moduli of the materials within the Earth behaved in such a simple way as in the examples we have been discussing so far, with everything quite constant in each layer and abrupt changes at interfaces. More likely, we might find the properties of the material changing more or less continuously with depth below the surface, and maybe find some discontinuities as well. It turns out that, inside the Earth, the velocity of seismic waves propagated downwards increases quite smoothly in most regions. In Section 4 we shall try to understand why this occurs but, first, what happens to the *direction* of wave propagation?

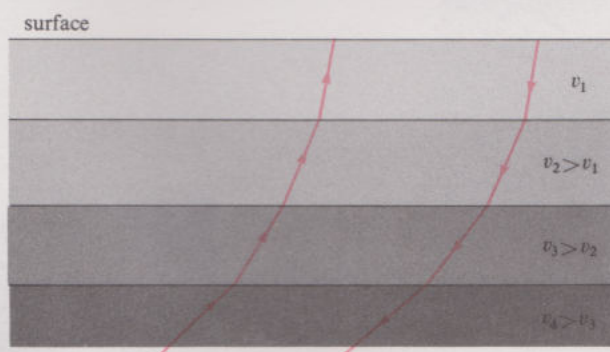


FIGURE 25 The propagation and refraction of seismic waves as velocity increases with depth.

Imagine that the medium consists of a series of layers, as in Figure 25, with velocity increasing each time the wave passes into the next layer down. At each interface the angle of refraction is greater than the angle of incidence. If, instead of increasing in jumps, the velocity increased continuously with depth as it passes through a series of very thin layers, then the propagation direction of a wave will follow a curved path (Figure 26). Eventually, the angle of incidence will be so large (greater than the critical angle) that the wave will be totally reflected. It now comes out towards the surface, going through a medium in which the velocity is progressively smaller, and the outgoing wave will follow a path that, like the ingoing one, is similarly curved. Thus, *if the velocity increases continuously with depth, the propagation path will be curved in some way such as in Figure 26. This is called continuous refraction.*

continuous refraction

Continuous refraction will occur for innumerable wave paths that we could describe radiating out from the source. Although we have shown only one wave in Figure 26, there are innumerable others, all curved, which reach a *maximum depth* of penetration half way between the source and wherever we choose to place our seismometer*. In the next Section you will see to what extent the Earth fits this simple model, where velocity increases smoothly with depth.

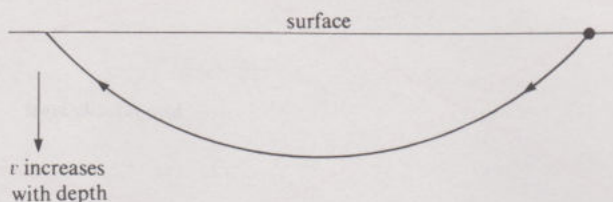


FIGURE 26 The path of a single seismic wave from an earthquake focus to a single seismometer in the case of velocity increasing smoothly with depth.

Now that you have completed Section 3 of Unit 4, you should be able to achieve the following Objectives, in addition to those listed earlier:

- Indicate your understanding of: angle of reflection, critical angle, total internal reflection, continuous refraction. (SAQs 13 and 14)
- State the law of reflection. (SAQ 13)
- Given the velocities of a wave in two adjacent media, predict the conditions under which total internal reflection will take place at the interface. (SAQ 14)
- Explain qualitatively why a seismic wave should follow a curved path if the propagation velocity increases smoothly with depth below the Earth's surface. (SAQs 13 and 16)
- As an extension of Objectives (b), p. 22 and (d), p. 26, given the relevant properties of two adjacent media, calculate the angle of refraction for any given angle of incidence and vice versa. (SAQ 15)

- SAQ 13** (a) In Figure 27, identify the angles of incidence, reflection and refraction.
 (b) In terms of the angles shown in Figure 27, state the laws of reflection and refraction.

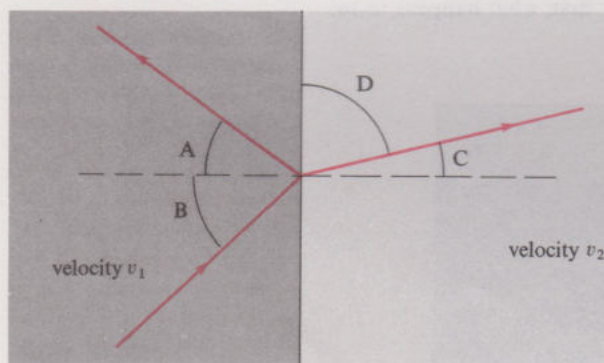


FIGURE 27 For use with SAQ 13.

SAQ 14 A wave passes from a medium A, where its velocity is v_A , to an adjacent medium B, where its velocity is v_B , as shown in Figure 28.

- Under what circumstances could this wave be totally reflected at the surface A/B?
- If $v_A = 0.5v_B$, for what angles of incidence will total reflection occur?

SAQ 15 If the density of medium A (Figure 28) is double that of medium B, and the axial modulus of medium A is two-thirds of that of medium B, what will be the angle of refraction of a P-wave passing from A to B if the angle of incidence is 30° ?

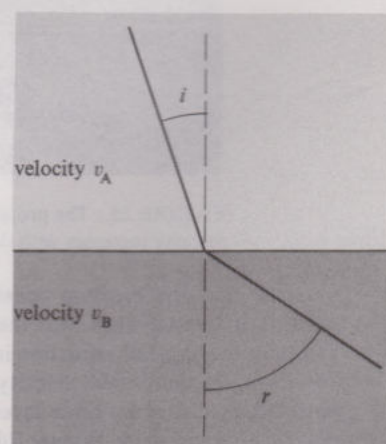


FIGURE 28 For use with SAQs 14 and 15.

* This actually assumes that the Earth is spherically symmetrical, which we shall demonstrate in the next Section.

SAQ 16 Under what circumstances could a seismic wave follow a path through the layers A and B of the form shown in Figure 29?

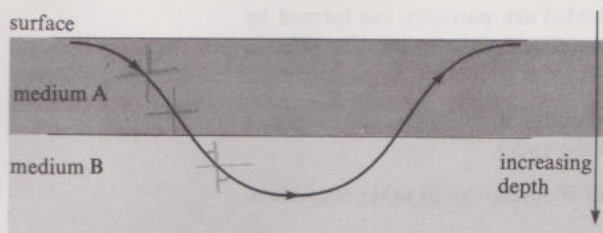


FIGURE 29 For use with SAQ 16.

4 The structure and internal composition of the Earth

Study comment In this Section, some of the rocks in your Home Experiment Kit are going to play an increasingly important part, and so, at this stage, we would like you to use the Audio-vision sequence: 'The origin of rocks'. This will take about one hour. Although we can establish the main features of the *internal structure* of the Earth in this Unit, a proper appreciation of the *internal composition* cannot be developed until later in the Course. In this respect, you should regard Section 4 as introductory, to be followed up in Units 27 and 28.

Collect together the items you need for using the Audio-vision sequence: your viewer, Filmstrips 4.1–4.3, Cassette AC90, and rock samples S1–S7. Lay out the samples in order in front of you and try to keep them in order during the exercise. You will need a penknife and the hand-lens from your Home Kit; and please make sure you have adequate light for examining rock specimens (e.g. a shaded reading-lamp). Next, find two sheets of A4 (Unit-sized) paper and write at the top of the first: IGNEOUS, and at the top of the second: SEDIMENTARY. Then on the left of the first sheet and equally spaced down the side write GRANITE, BASALT, PERIDOTITE, GABBRO; and similarly, on the second sheet: SANDSTONE, LIMESTONE.

When you have done this, place the paper, face down in a convenient place—you will use it near the end of the tape sequence.

Before you listen to the tape, which takes 30 minutes to run straight through, we advise you to spend 10 minutes examining specimens S1–S7, noting their features and particularly their differences. When you are ready, with the first filmstrip (4.1) in the viewer, switch on the cassette machine.

When you have completed the Audio-vision sequence, and before reading the rest of Section 4, check that you can achieve the following Objectives. You should be able to:

(a) Indicate that you understand the following terms: igneous rock, magma, sedimentary rock, granite, basalt, gabbro, peridotite, sandstone, extrusive and intrusive rocks, effusive and explosive volcanism. (SAQ 17)

(b) Explain how the textures and grain size of different rocks provides useful information about the processes involved in their formation. (SAQ 17)

igneous rock magma sedimentary rock
granite basalt gabbro peridotite
sandstone extrusive rock intrusive rock
effusive volcanism explosive volcanism

SAQ 17 Which of the following statements are true and which are false?

- (a) Igneous rocks usually are formed by the crystallization of molten lava or magma.
- (b) Sedimentary rocks crystallize from melts at temperatures below 100 °C.
- (c) Granite, basalt and peridotite are all examples of extrusive igneous rocks.
- (d) Basalt is a relatively fine-grained igneous rock that cooled from a melt relatively quickly compared with granite.

- (e) The grain size and texture of most granites indicates that they crystallized in intrusions below the Earth's surface.
- (f) Some igneous rocks, composed of welded ash particles, are formed by effusive volcanism, which frequently is associated with the emission of large volumes of gas.
- (g) Sandstone is a sedimentary rock composed of fragments or grains derived from the weathering of pre-existing rocks.
- (h) Gabbro is a coarse-grained rock that is similar in all other respects to fine-grained basalt.

4.1 The density of Earth materials

It is about time we started to apply your observations to the Earth's interior! First compare the density of your rock samples with that of the Earth. Specimen S1 is granite, an extremely common rock, widely distributed on all continents, which you will also find in bank facades, gravestones and memorials—including Cleopatra's needle. Your Home Kit specimen came from Peterhead in Aberdeenshire.

Granite is one of the most abundant rocks at the surface of the Earth. Could the entire Earth be made of granite?

Recall from Section 1.2 that the average density of the Earth is $5.5 \times 10^3 \text{ kg m}^{-3}$. Granite has a density of about $2.6 \times 10^3 \text{ kg m}^{-3}$, less than half that of the Earth. Unless you could find some way of squashing up the minerals of granite to more than double their densities, it seems that the entire Earth is not made of granite.

So we need something heavier than granite inside the Earth—sandstone will not do because it is not sufficiently dense and, by definition, sediments occur only near the surface.

What about the two other rocks whose density you determined?

Samples S3 and S4, basalt and peridotite, have densities of $2.9 \times 10^3 \text{ kg m}^{-3}$ and $3.3 \times 10^3 \text{ kg m}^{-3}$, which, although they represent an increase in density over granite, are still not high enough to account for the much larger density of the whole Earth.

You may recall that our most dense rock, peridotite, occurs only where material has been uplifted from 'deep' within the Earth. Although we are looking at material from perhaps a few tens of kilometres depth, there is a trend towards rocks of higher density coming from deeper levels in the Earth and this is what we might expect. So what can we conclude about the deeper interior of the Earth? Clearly, there must be something denser inside than peridotite, basalt and granite in the form we find them at the surface. Now there are two ways in which density might increase with depth in the Earth:

1 The density might increase owing to the *compression* of the deeply buried rocks by the weight of those overlying. Think about what happens when a car goes to the scrap yard and is compressed into a cube, just a few feet in dimensions. Clearly the total volume it occupies is reduced, but the mass is unchanged, so the density must increase. In the same way, the extremely small particles that go to make the minerals that make the rocks will be forced closer and closer together the deeper you go inside the Earth and the density will progressively increase.

2 On the other hand, the density might increase inside the Earth owing to a change in the *chemical composition* of the material. For example, if you took a cube of lead of the same volume as the compressed car, you would find that the mass, and therefore the density, would be greater than that of the largely iron car, even though the two volumes would be the same.

This raises the possibility that the Earth might have a dense, metallic core. But stop and think. What reasons have you for supposing that the Earth has a *core*—other than the recollection of reading about it somewhere? Why shouldn't the dense material be distributed like raisins in a plum pudding? In short, what reasons have you for choosing A rather than B in Figure 30? Both satisfy the density conditions perfectly.

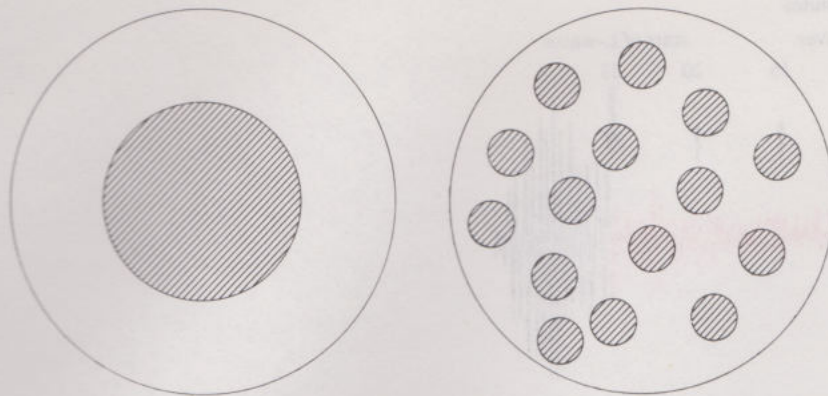


FIGURE 30 Possible distributions of high-density (shaded) material within a planetary body such as the Earth: (A) the core model; (B) the raisin model. Say, for simplicity, that each has one-quarter of its volume with $\rho = 13 \times 10^3 \text{ kg m}^{-3}$ and three-quarters of its volume with $\rho = 3 \times 10^3 \text{ kg m}^{-3}$. Then both models have an average density of $5.5 \times 10^3 \text{ kg m}^{-3}$. How can we tell which kind of core the Earth has?

By this stage, it must be pretty obvious that the answer comes from seismic waves. Most people have seen the effective way in which X-rays can be transmitted through a human body to reveal the internal skeleton. In the same way, seismic waves are used to determine the internal structure of the Earth because they travel *right through the Earth's interior*. Seismologists exploit the dependence of seismic wave velocities on density and also elastic moduli (Section 3.1), which in turn leads to detectable reflection and refraction layers within the Earth (Sections 3.2 and 3.3). To locate such layers, we need to know certain things about the earthquake event and the recordings made elsewhere on the Earth's surface.

4.2 Seismic records

Suppose that an earthquake occurs at some point near the Earth's surface, and that both its time and the position of its focus, and hence epicentre, are known. At a recording station somewhere on the Earth's surface, the arrivals of the various waves from the earthquake can be timed. When using these data, the distance between the epicentre and recording station is described by the *epicentral angle*.

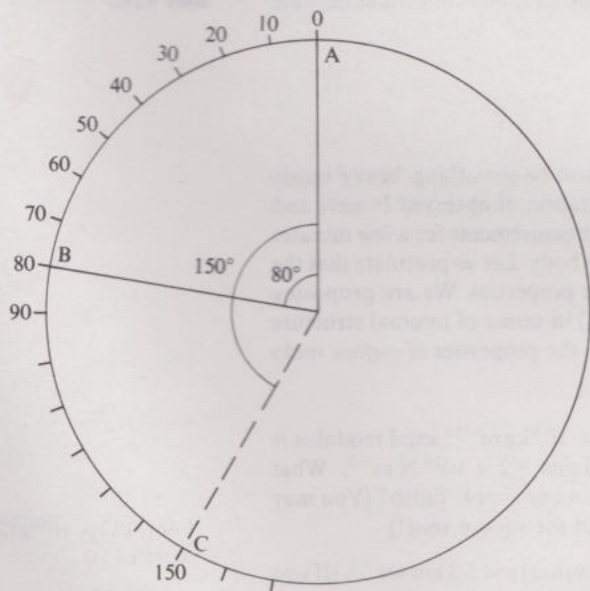


FIGURE 31 The relationship between distance over the Earth's surface between an earthquake at A and recording stations at B and C is expressed as epicentral angles of 80° and 150° respectively. Since the circumference of the Earth is about 40 000 km, every 10° epicentral angle corresponds to about 1 100 km between epicentre and recorder. ($40\,000 \text{ km per } 360^\circ = 110 \text{ km per degree}$).

This is the angular difference between them at the centre of the Earth. In Figure 31 you can see that every 10° corresponds to roughly 1 100 km surface distance between epicentre and recording station.

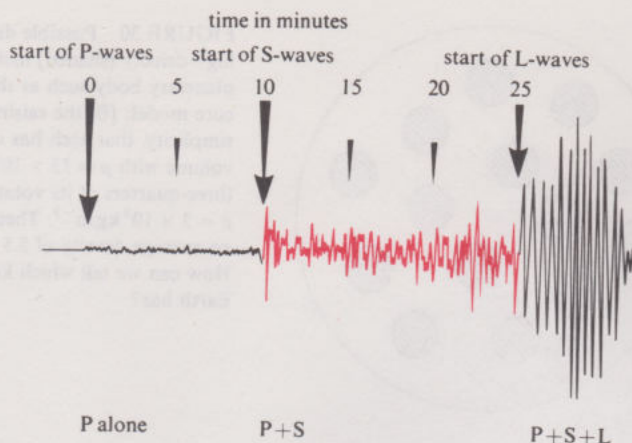


FIGURE 32 Seismogram of an earthquake with epicentre at Erzincan, Turkey, recorded at Cambridge, Massachusetts, USA, on 26 December 1939. The first arrivals (left) are small deflections made by P-waves alone; then come large deflections made mainly by S-waves, with a small P contribution; and finally there are even larger deflections mainly from L-waves, but with small P and S contributions. The time that elapsed between the start of the P-waves and the start of the S-waves indicates that the distance from epicentre to recording station is about 77° or 8550 km.

Figure 32 shows a typical seismogram of an earthquake recorded at an epicentral angle of 77° (8550 km) from the epicentre. First, it shows small deflections, and these are the P-waves, which actually arrived about 11 minutes after the earthquake event. Recalling from ITQ 3 that the average velocity of P-waves in the solid material of the Earth is almost twice that of S-waves, it should not surprise you to see the S-wave pulse arrive 10 minutes after that of the first P-waves (Figure 32). The very much larger deflections, which appear last on the seismogram, are due to waves which travel around the *surface* of the Earth; the letter L is used to designate them. This type of wave is due to the 'rippling' or vibration of the entire surface, where the Earth is relatively unconstrained because it is in contact with the atmosphere. Although these waves travel slowly they are the earthquake waves most like water waves. But here, where we are interested in the Earth's internal structure, it is the *body waves* (P and S) which concern us because they travel through the Earth.

surface waves (L-waves)

body waves

4.3 A simple seismic model

Although we know from density data that there must be something 'heavy' inside the Earth, to help you to understand the interpretation of observed P-wave and S-wave travel times, we shall forget about density measurements for a few minutes and consider wave propagation in a homogeneous body. Let us postulate that the Earth is completely *uniform* in density and elastic properties. We are proposing the simplest possible Earth *model* (see Section 1.1) in terms of internal structure and composition. In this model let us assume that the properties of *surface rocks* apply right through the Earth.

An average density for surface rocks is $2.9 \times 10^3 \text{ kg m}^{-3}$; axial modulus is about $9.2 \times 10^{10} \text{ N m}^{-2}$ and rigidity modulus $3.2 \times 10^{10} \text{ N m}^{-2}$. What P-wave and S-wave velocities should apply to our simple Earth? (You may use your calculator for this, and don't forget the square root!)

P-wave velocity is 5.6 km sec^{-1} and S-wave velocity is 3.3 km sec^{-1} . (If you had any difficulty here, refer back to equations 16 and 18 of Section 3.1.)

For $\sqrt{(9.2 \times 10^{10})/(2.9 \times 10^3)}$, the keying sequence is:

9.2, EE, 10, $\div$, 2.9, EE, 3, =, $\sqrt{x}$

ITQ 6 If the velocities of P-waves and S-waves are constant in the Earth, how long would it take for first arrivals to occur at epicentral angles of 180° from a near-surface earthquake? (Assume an Earth diameter of 12740 km: Figure 1.)

Now, in such a hypothetical uniform Earth, there can be no reflections and refractions, and so earthquake waves travel in *straight lines*. The first arrival times will be simply proportional to the shortest straight-line distance through the Earth from the source to the recording station. Repeating ITQ 6 for the entire surface of the Earth, we obtain the results shown in Figure 33. It is possible to

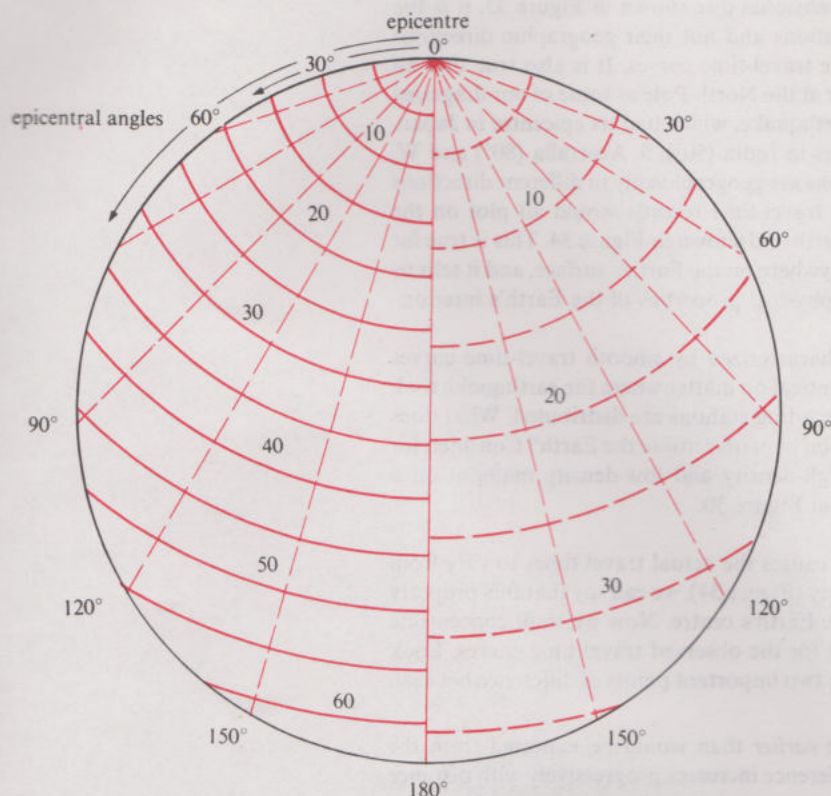


FIGURE 33 Section through a hypothetical Earth in which v_p and v_s are constant at 5.6 and 3.3 km s^{-1} . Travel paths are shown (dashed) to stations at 30° epicentral angle intervals from the source and the heavy lines indicate how far P-waves (right) and S-waves (left) will travel a given time (minutes).

synthesize the information in Figure 33 by drawing *travel-time curves*, where travel time is plotted against epicentral angle for P-waves and S-waves. These travel times then can be compared with those we actually observe from natural earthquakes to find out whether earthquake wave velocities are greater or less than predicted by our uniform, homogeneous Earth model. Figure 34 makes just this comparison and, not surprisingly, you can see that the Earth is not simply homogeneous! How should this simple model be modified?

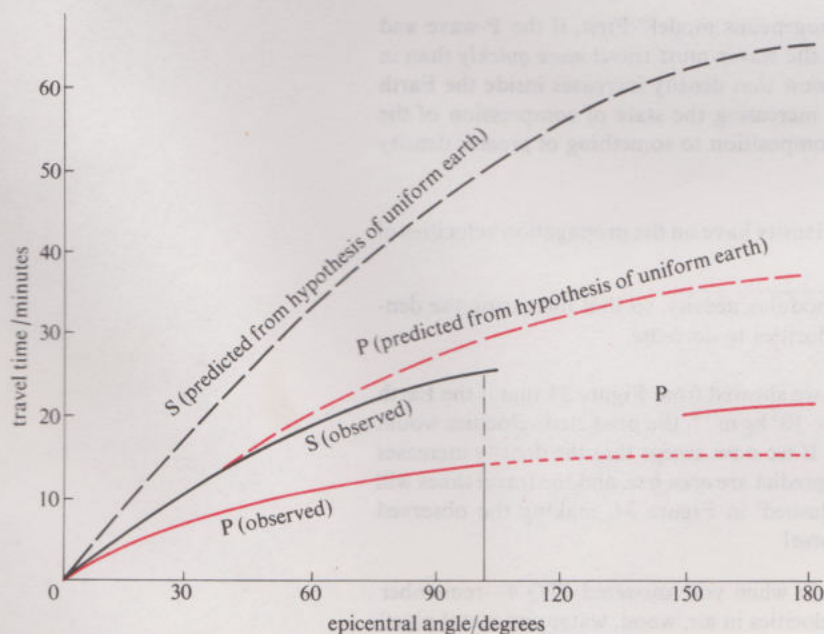


FIGURE 34 Travel-time curves for P-waves and S-waves in a hypothetical Earth of uniform density and elastic moduli such that $v_p = 5.6$ and $v_s = 3.3 \text{ km s}^{-1}$ (dashed curves), compared with observations for the real Earth (solid curves). The curve with short red dashes is the theoretical extension of the observed travel-time curve for P-waves (see p. 38).

The observed P-wave travel times between 150° and 180° are longer than predicted on the basis of those at low epicentral angles, and the explanation for this lower velocity appears in Sections 4.4 and 4.5.

4.4 Modifying the simple model

It is quite a straightforward, but time-consuming, business to calculate travel-time curves based on the evidence of an earthquake, of known epicentre, that has been recorded at various places over the Earth's surface. Although the real Earth is almost spherical and not the two-dimensional disc shown in Figure 33, it is the distance to the different recording stations and not their geographic directions which is important in determining the travel-time curves. It is also true that all earthquakes do not conveniently occur at the North Pole as some of our diagrams might suggest! Say we recorded an earthquake, which had its epicentre in Japan, at stations of varying epicentral angles in India (50°), S. Australia (80°) and W. Canada (90°). Although all these stations are geographically in different directions from the epicentre, nevertheless their travel-time records would all plot on the smooth curves observed for the real Earth and shown in Figure 34. This is true for any set of records of an earthquake anywhere on the Earth's surface, and it tells us something very important about the physical properties of the Earth's interior.

ITQ 7 All earthquakes are characterized by smooth travel-time curves (Figure 34) that are virtually identical no matter where the earthquake took place and no matter how the recording stations are distributed. What does this tell you about the distribution of matter inside the Earth? Consider, for example, the distribution of high-density and low-density material on a core or raisin model as shown in Figure 30.

So, whatever property of the interior causes the actual travel times to vary from those predicted by a homogeneous body (Figure 34), we can say that this property is symmetrically distributed about the Earth's centre. Now we shall concentrate on this 'property' and try to account for the observed travel-time curves. Look again at Figure 34. You should notice two important points of difference between the observed and predicted curves:

- 1 Both P-waves and S-waves arrive *earlier* than would be expected from the homogeneous model, and the time difference increases progressively with distance from the epicentre. This shows that P-waves and S-waves are travelling faster (i.e. with higher velocity) than we would predict from the homogeneous model.
- 2 Beyond a distance represented by an epicentral angle of 103°, predictions from the homogeneous model and actual observations do not resemble each other even remotely. No traces of S-waves are recorded and a *gap* appears in the P-wave travel-time curve.

4.4.1 Explaining the observed velocities

How can we modify the simple homogeneous model? First, if the P-wave and S-wave travel times are shorter, then the waves must travel *more quickly* than in the homogeneous model. Now we know that density increases inside the Earth and this can be due to two causes: increasing the state of compression of the material or changing the chemical composition to something of greater density (Section 4.1).

What effect will increasing the density have on the propagation velocities of seismic waves?

You know that $\text{velocity} = \sqrt{\text{modulus}/\text{density}}$, so that increasing the density will cause seismic wave velocities to *decrease*.

But there is something wrong here—we showed from Figure 34 that if the Earth had the density of surface rocks: $2.9 \times 10^3 \text{ kg m}^{-3}$, the predicted velocities would be less than those actually observed. If we now accept that the density increases with depth, the new velocities we can predict are *even less*, and the travel times will become even greater than shown 'dashed' in Figure 34, making the observed predicted travel-time discrepancy worse!

You have already solved this problem when you answered ITQ 4—remember comparing the densities and sound velocities in air, wood, water and metal pipe? Sound velocities actually increased with density, so we concluded that elastic modulus increases faster than density. The same conclusion applies here: if we postulate that, *in the Earth*, elastic moduli increase with depth *more rapidly than does the density*, then seismic velocities of the homogeneous model will increase,

resulting in a better match between observed and predicted travel times. Physically, density probably increases because of increasing compression of the material by the weight of overlying rocks (Section 4.1). As this occurs, the 'balls and springs' are squashed together: not only does density increase but the material becomes more rigid and less compressible. In this way, the rigidity and axial moduli appear to increase more rapidly than density. It is thought that this applies to waves emerging up to epicentral angles of 103° . This is because the time discrepancy in Figure 34 increases towards 103° —this is explained by the dominant increase of elastic moduli with depth.

In this medium, in which velocity increases continuously with depth, what will happen to the *direction* of wave propagation?

Recall from Section 3.3.3, and particularly Figures 25 and 26, that increasing velocity will result in *continuous refraction* away from the vertical until the critical angle is reached and the wave starts to return to the surface. The wave will be curved, concave with respect to the Earth's surface.

Continuous refraction of a curved wave in the Earth's interior is illustrated in Figure 35, and it follows that the greater distance between source (S) and receiver (R), the deeper the wave has penetrated.

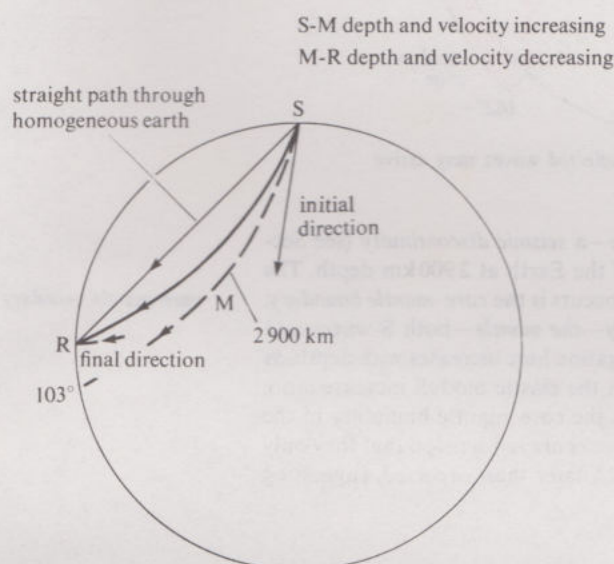


FIGURE 35 Continuous refraction of an earthquake wave between source (S) and receiver (R), also showing that an epicentral angle of 103° corresponds to 2900 km depth of penetration in the Earth.

So far, our modified model is no longer that of a *uniform* Earth, but of a *continuous* one with no sudden *breaks* or *discontinuities* in its structure. We now have a picture of a medium where wave velocities increase continuously with depth because, under increasing compression, ψ and μ increase faster than ρ .

4.4.2 What happens beyond 103° degrees?

We can determine the way in which elastic moduli and density vary with depth (see the values in Table 2 at the end of Section 4) by altering them until the predicted travel times match the observed ones, until everything goes wrong at 103° . Apparently some drastic change occurs in the properties of the Earth's interior at a particular depth, and it has been calculated that waves reaching stations just inside the 103° limit have penetrated the Earth to a maximum depth of 2900 km—see Figure 35. There are three features of the travel-time curves *beyond* 103° to explain (see Figure 34):

- 1 No direct S-waves are recorded at distances from the epicentre corresponding to angles greater than 103° .
- 2 Between 103° and 142° no direct waves of any kind are received.
- 3 Between 142° and 180° direct P-waves are once more received, but if we continue the existing travel-time curve for epicentral angles less than 103° up to

142° (short red dashes in Figure 34), we find that the actual travel times are longer than predicted. The region between 103° and 142° is called a *P-wave shadow zone* because no direct waves are recorded.

Incidentally, we used the term *direct* waves in 1 to 3 above because some P-waves are recorded in the shadow zone, but these are *indirect*, having been reflected from the Earth's surface as shown in Figure 36. They can be identified by their longer arrival times, due to their longer travel paths. Only *direct* waves are shown in the travel-time diagram (Figure 34).

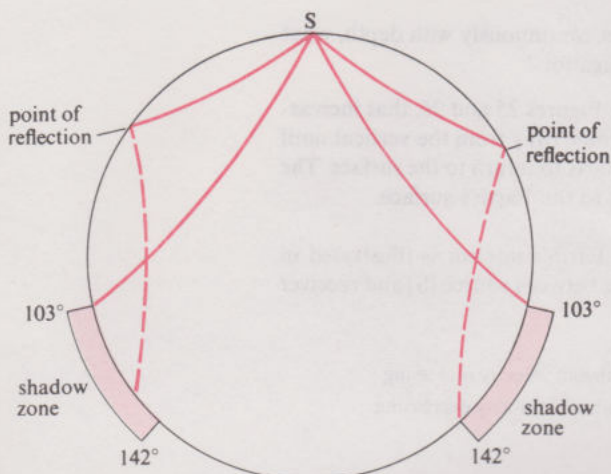


FIGURE 36 Diagram showing that *reflected* waves may arrive between 103° and 142°.

There is evidently a sudden qualitative change—a *seismic discontinuity* (see Section 3.2)—in the composition and structure of the Earth at 2900 km depth. The boundary surface at which this sudden change occurs is the *core–mantle boundary*. In the part of the Earth outside this boundary—the *mantle*—both S-waves and P-waves are propagated. The velocity of propagation here increases with depth, as do the density and the elastic moduli—though the elastic moduli increase more rapidly than density. But what happens across the core–mantle boundary in the Earth's core? S-waves no longer occur and *P-waves are refracted* so that they only reappear at epicentral angles greater than 142°, later than expected, suggesting that they have travelled more slowly.

4.5 The Earth's core

S-waves stop at the core–mantle boundary. What does this suggest to you about the core? (Refer back to the end of Section 3.1 if you have difficulty.)

Shear waves travel through solids of all types but cannot be transmitted through liquid. So it seems reasonable to infer that the core has the properties of a liquid in that its resistance to shear is effectively zero. There is no need, of course, for all the core to be liquid—if only the outer part were liquid it would stop the propagation of the S-waves.

We know from the disappearance of S-waves that the outer part of the core is liquid. How about P-waves and, in particular, the P-wave *shadow zone* between epicentral angles of 103° and 142° (Figure 36). Why are there no P-waves in this zone?

If the core has liquid properties, you would still expect it to transmit P-waves. Can you suggest what might happen to P-waves which penetrate the Earth just below 2900 km?

First, if v_p increases across this boundary, there will be a complex region of multiple or duplicated arrivals at epicentral angles $\leq 103^\circ$ as P-waves are *reflected* from the core at the critical or greater angles of incidence. We do not see these multiple arrivals. So v_p must decrease across this seismic discontinuity and P-waves are *refracted towards* the normal as they strike the interface (see Section 3.2).

4.5.1 The core–mantle boundary

So P-wave velocities drop as they enter the core, and this explains why—between 142° and 180° epicentral angles—their first arrivals are recorded later than expected when we extend the mantle travel-time curve (Figure 34). But why are P-wave velocities less in the outermost parts of the core than in the mantle?

Recall that density and elastic moduli both increase with depth in the mantle, the latter more rapidly than the former. What changes in these physical properties can occur across the core–mantle boundary? Consider simultaneous changes in *both*.

Remembering that velocity decreases and that $v_p = \sqrt{\frac{\psi}{\rho}}$:

- (a) density increasing with axial modulus constant or decreasing slightly;
- (b) a density decrease outweighed by a large decrease in axial modulus.

Because we are expecting to find a high-density core inside Earth (Section 4.1) and because the axial modulus (resistance to compression) of the *fluid* core is likely to be less than if it were solid (cf. ice and water—answer to SAQ 10), alternative (a) is by far the more sensible deduction. It is now accepted by seismologists that density increases *and* axial modulus decreases downwards across the core–mantle boundary. But by how much? Let us make the assumption that all the mantle is *peridotite* (Section 4.6.2). Because of compression (which will increase the density of peridotite compared with that at the surface), the density at the base of the mantle has been calculated to be $5.4 \times 10^3 \text{ kg m}^{-3}$. How about the outer core? Well, there is good reason to believe (Section 4.5.3) that an *iron–sulphur mixture* is present there, with a probable density at that depth of $9.9 \times 10^3 \text{ kg m}^{-3}$. This is a huge change of density: no wonder P-wave velocities drop! Axial modulus values have also been calculated for peridotite ($10.0 \times 10^{11} \text{ N m}^{-2}$) and for iron–sulphur liquid ($6.5 \times 10^{11} \text{ N m}^{-2}$). We now have sufficient information to calculate the P-wave velocities on either side of the core–mantle boundary: $v_p = 13.6 \text{ km s}^{-1}$ for the base of the mantle, and $v_p = 8.1 \text{ km s}^{-1}$ in the outermost core. As you will see, these values allow us to predict the amount of refraction that will occur across the boundary. The predictions compare favourably with actual observations and give us confidence in the values of ψ , ρ , and v_p that we used to make them.

Now you can calculate the amount of refraction that will occur when a P-wave just grazes and enters the core (see Figure 37). The angle of incidence will be 90° : what is the angle of refraction?

From Snell's law:

$$\frac{\sin i}{\sin r} = \frac{\sin 90^\circ}{\sin r} = \frac{1}{\sin r} = \frac{v_{\text{mantle}}}{v_{\text{core}}}$$

So $\sin r = 8.1/13.6$

Therefore $r = 36\frac{1}{2}^\circ$

(Notice that, by analogy with Figure 23, this angle r is actually a *critical* angle of refraction.)

You have now found that the first wave which just enters the core at 2900 km depth is bent *strongly*—by $53\frac{1}{2}^\circ$ (Figure 37).

By how much will waves that hit the core–mantle boundary at lower angles of incidence be refracted? Consider, at least, cases where $i = 42^\circ$ and $i = 0^\circ$.

From Snell's law we know that:

$$\frac{\sin i}{\sin r} = \frac{13.6}{8.1} = 1.68$$

Consequently, as i decreases, so does r , but less rapidly until when $i = 0$, $r = 0$ and we have an undeviated wave. In the case where $i = 42^\circ$, then $r = 23\frac{1}{2}^\circ$ and the wave is bent by $18\frac{1}{2}^\circ$, much less than for the 'critical wave'.

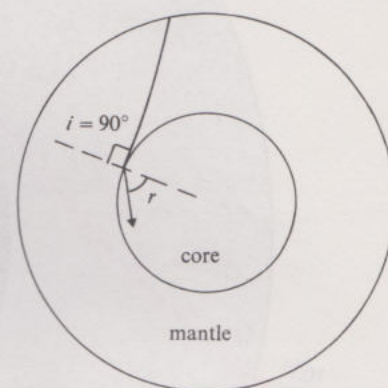


FIGURE 37 For use in calculating r when $i = 90^\circ$: a core-grazing wave.

Calculator keying sequence: given $\sin r = 8.1/13.6$, to find r in degrees:

$8.1, \div, 13.6, =, \text{INV}, \sin$

If you get a wrong answer, check that your calculator is in the degree mode.

4.5.2 Seismic waves inside the core

The next step is to see what happens to these waves as they pass through the core. In Figure 38 we show *curved* wave paths away from the points on the core surface B and E that you have just considered—why? You will remember that continuous refraction caused wave paths to be curved in the mantle because of progressive increases in propagation velocities. As waves that pass through the core appear earlier than they would if v_p was 8.1 km s^{-1} right across the core, the velocity must increase with depth, and v_p of 10 km s^{-1} at 5000 km depth has been calculated. So the wave paths are curved and, like the solid mantle, the axial modulus of the fluid core medium is increasing more rapidly than density as the material is compressed.

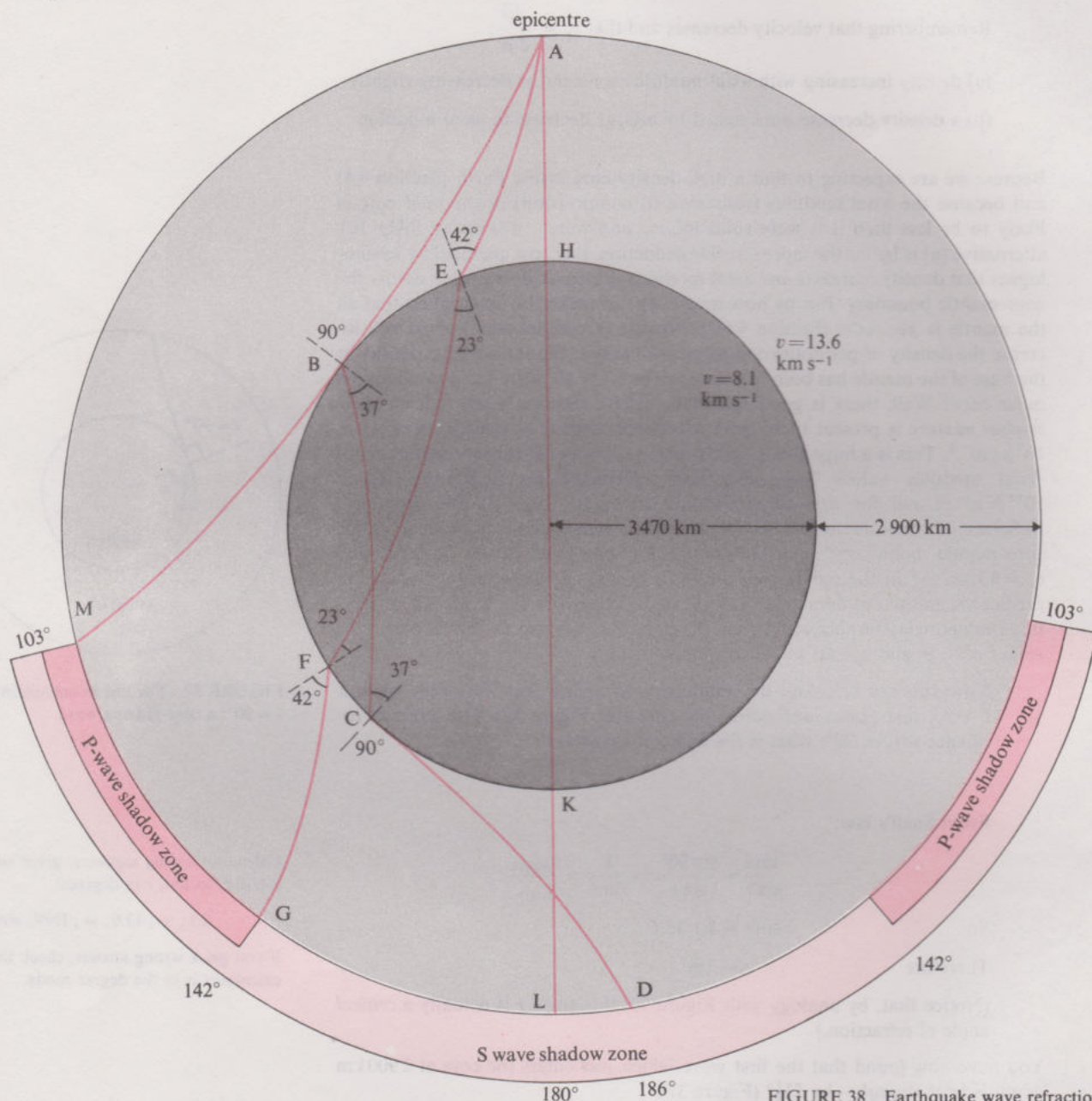


FIGURE 38 Earthquake wave refractions at the core-mantle boundary showing the origin of the P-wave shadow zone after transmission across the core. Letters refer to particular waves described in the text. Angles are quoted to the nearest whole number.

So the waves AB and AE (Figure 38) follow the *symmetrical curved paths* BC and EF such that the angles of refraction at B and E become the angles of incidence at C and F. The waves are now bent back away from the vertical by the same amount as they were originally refracted. But the amount by which the waves have already been deviated means that there are no arrivals between 103° and 142° —hence the *P-wave shadow zone*. This is because waves between AB and AE eventually emerge between 186° (D) and 142° (G) from the epicentre (Figure 38), but the waves between AE and AH move back round from 142° (G) to 180° (L). The wave AHKL hits every boundary at an incident angle of 0° (vertical) and so is undeviated.

The following few paragraphs of text describe the seismological evidence used to establish the existence of a *solid inner core* of radius 1215 km at the centre of the Earth, some of which we shall demonstrate on TV. You should regard it very much as optional reading; it is essential only that you know of the existence of the inner core before leaving Section 4.5.2. As before, the optional section starts with a star ★ in the margin and ends with a double star ★★.

So far we have assumed that the whole core is a fairly uniform fluid in which v_p increases with depth only because of compression. If we look at the waves emerging near L in Figure 38 that belong to the 'vertical' group near AHKL, it turns out that they arrive about 25 seconds *earlier* than predicted by the uniform model. Remember that these waves are the only ones to travel close to the centre of the Earth.

ITQ 8 Can you think of any reasons why v_p might be higher than expected close to the centre of the Earth?

There are two recently discovered seismological facts that settle this issue. Look at Figure 39, which shows a few more wave paths.

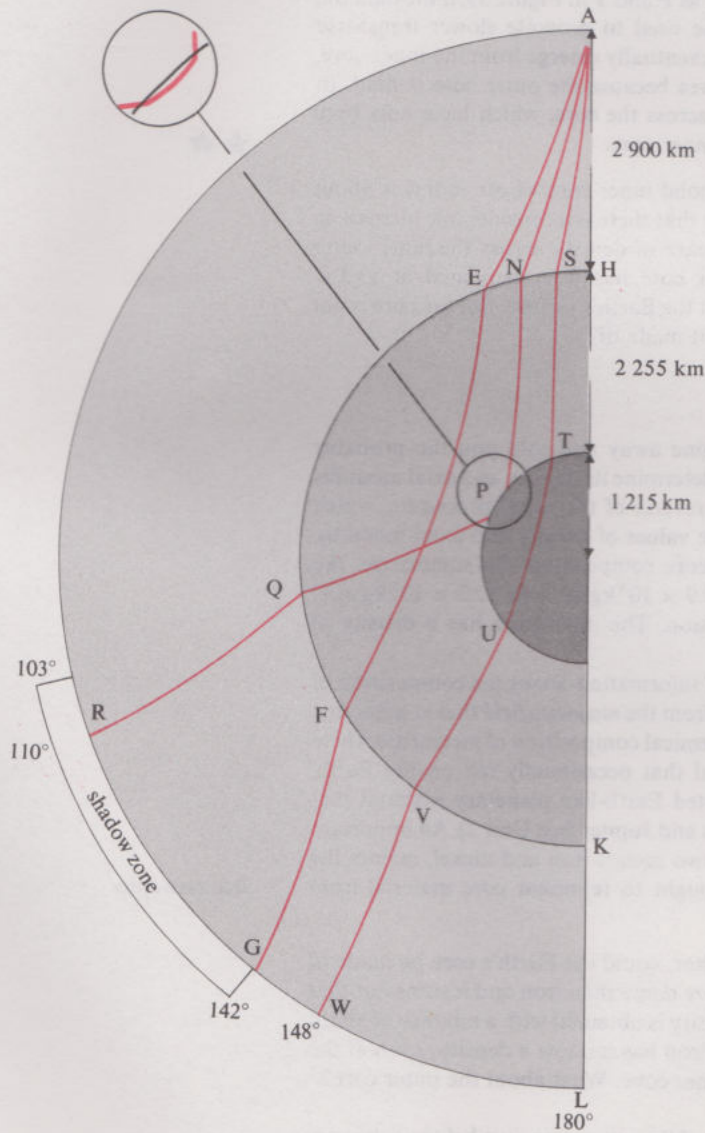


FIGURE 39 Earthquake wave refractions caused by the inner core. An outer core wave (AEFG) as in Figure 38 is shown for reference. (Insert shows an enlargement of the point P.)

Starting with a near surface focus at A and considering waves that have smaller angles of incidence on the core-mantle boundary than AE (for which $i = 42^\circ$), we reach a wave AN which is known to emerge at an epicentral angle of 110° —right in the middle of the shadow zone! This wave is strongly deviated by another seismic discontinuity at 5155 km depth, this time *away* from the normal to this interface. It took a long time for the wave ANPQR to be unscrambled by seismologists from the many reflected arrivals in the 'shadow zone' but, once found, it

proved the existence of an *inner core*, radius 1215 km, surrounded by an *outer core*. What is more, the strong refraction of the wave at P indicates that the inner core has the *higher velocity*, so if you reconsider the wave AHKL, which traverses the entire diameter of the inner core, there is no wonder it arrives 25 seconds early!

inner core
outer core

The deepest parts of the outer core have P-wave velocities of about 10.1 km s^{-1} . Estimate the average v_p for the inner core, using the distance from Figure 39.

The wave must traverse 2430 km which at 10.1 km s^{-1} would give a travel time of 241 seconds. Now, to arrive 25 seconds earlier; that is in 216 seconds, v_p must be 11.2 km s^{-1} . So v_p is about 1 km s^{-1} higher inside the inner core, and the evidence of strong refractions, such as for wave ANPQR, tells us about the existence and size of the inner core. What else do we know about it?

It has been found that yet another set of arrivals occurs near L, this time about 3 minutes *later* than the first set. This set of arrivals had been predicted as a test for the possible *solid* nature of the inner core. What was anticipated is that when a P-wave strikes the inner core at places such as P and T in Figure 39, if the material is solid, *part* of the wave's energy must be used to generate slower transverse S-waves across the inner core. When they eventually emerge from the inner core, these S-waves must regenerate only P-waves because the outer core is fluid. In other words, the passage of P-S-P waves across the core, which have now been found, testifies to the solid nature of the inner core.



So, at the centre of the Earth, there is a solid inner core whose radius is about 1215 km, as shown in Figure 39. We know that there is a considerable increase in axial modulus, coupled with a slight increase of density across the inner-outer core boundary. The density of the inner core has been estimated at $13.5 \times 10^3 \text{ kg m}^{-3}$, about the density of mercury at the Earth's surface. But the core is not thought to be made of mercury! What is it made of?

4.5.3 The composition of the core

In Section 4.5.1 we actually gave the game away and told you the probable composition of the outer core in order to determine its density and axial modulus and hence its seismic velocity. This is the reverse of the usual procedure, which is to use seismic data to determine possible values of density and axial modulus. Density is the most useful constraint on core composition. To summarize, the outer core has a density variation from $9.9 \times 10^3 \text{ kg m}^{-3}$ to $12.3 \times 10^3 \text{ kg m}^{-3}$ with depth, owing to increasing compression. The inner core has a density of $13.5 \times 10^3 \text{ kg m}^{-3}$.

Density values are an important source of information about the composition of the core, but to these we can add evidence from the *magnetic field* that is generated in the outer core (Unit 5) and from the chemical composition of meteorites: these are fragments of extra-terrestrial material that occasionally fall on the Earth. Meteorites are probably pieces of disrupted Earth-like planetary material that originate in the asteroid belt between Mars and Jupiter (see Unit 2). An important group are composed of a mixture of the two metals iron and nickel, mainly the former, and these *iron meteorites* are thought to represent core material from broken-up planets (cf. Unit 28).

iron meteorites

Adding these pieces of information together, could the Earth's core be made of iron and nickel? Well, *nickel* is a little *more dense* than iron and it turns out that a good 'fit' to the observed inner core density is obtained with a mixture of about *40 per cent nickel, 60 per cent iron*. Pure iron has too low a density, even at the estimated state of compression for the inner core. What about the outer core?

ITQ 9 At the compressional state of the outer core, just below its boundary with the mantle, pure liquid iron is known to have a density around $10.6 \times 10^3 \text{ kg m}^{-3}$. On this basis, must the outer core be made of iron and nickel or something else?

On various geochemical grounds as well as density it is thought that *sulphur* is combined with iron in the outer core, forming the compound *iron sulphide*. Similar material is found at the Earth's surface in ore deposits, where it is recognized

as a yellowish metallic-looking substance, often termed 'fool's gold'—not without some justification! The mineral name given to iron sulphide is *pyrite*. In fact, meteoritic iron sulphide contains less sulphur and more iron than pyrite, and the same is probably true of the Earth's outer core—the detailed arguments need not worry us here. Liquid iron sulphide is an ideal medium in which to generate the Earth's *magnetic field*, as you will find in Unit 5. But as to *why* the outer core is *liquid* and the inner core *solid*, we must ask you to be patient until Unit 28!

4.6 The Earth's crust and mantle

However attractive it is to speculate about the Earth's core composition, there is little hope that we shall ever obtain actual samples of the material from such great depths. There are other regions inside the Earth which are more accessible: the *crust* and *mantle*. In Section 4.4, the mantle was defined seismologically as comprising the region penetrated by P-waves that emerge at up to 103° epicentral angles. It seems from the travel-time diagram (Figure 34) as though there are no complications and we might infer that smooth increases of velocity, density, axial and rigidity moduli occur with depth. All this is true to a first approximation but, in detail, remember that there are different rock types to fit into the picture. Some of these you have examined—granite and peridotite, for example—and all of them have different densities. You might also ask why we want to introduce complications by having a *crust* as well as a *mantle*.

The story dates from the early twentieth century when a Yugoslavian seismologist, Mohorovičić (čič is pronounced 'chic') studied seismograms for shallow-focus (< 40 km) continental earthquakes recorded over fairly short distances (< 800 km) from the epicentre. The surprising result was that two very distinct sets of P-waves and S-waves were recorded.

How do you think this could happen? Think again about refractions within the Earth.

Mohorovičić deduced that there must be two distinct paths involved: one set of waves travelled directly, but the second set must have been strongly refracted after they had started moving downwards into the Earth.

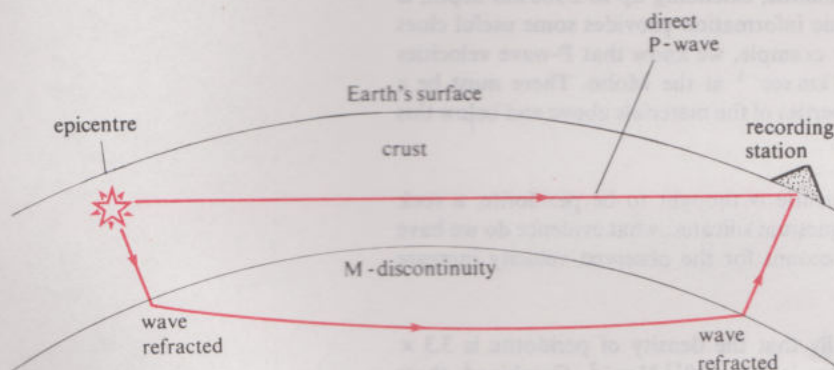


FIGURE 40 The Mohorovičić discontinuity, showing how both a direct crustal P-wave and a refracted, mantle P-wave will be recorded for focus-recorder distances of several hundred kilometres.

Mohorovičić's idea is illustrated in Figure 40, where the first, or direct, wave-set passes through an upper region of fairly uniform low velocity, whereas the second set travels downwards, is strongly refracted by a higher velocity medium, and is refracted back to the surface because of the curvature of both the velocity boundary (parallel to the surface) and the wave paths.

Does this remind you of refractions anywhere else in the Earth? What do we call this sort of velocity boundary?

This is a seismic discontinuity, just like the core–mantle boundary.

This particular discontinuity is present all around the Earth and separates the *crust*, where v_p is between 6 and 7 km s^{-1} from the *mantle*, where v_p starts at 8 km s^{-1} . It has been named the *Mohorovičić discontinuity*, or *Moho* for short, and always occurs within 90 km of the Earth's surface. Finally, note that our modern knowledge of crustal structure is based almost entirely on explosion seismology, in which geophysicists choose where the shot should be fired, rather than waiting for a natural earthquake!

Mohorovičić discontinuity

4.6.1 The composition of the crust

Although we live on the Earth's crust and can sample most of it at the surface or by drilling holes, we still have to rely on seismic data to determine how thick it is. In fact, the thickness of the crust varies considerably from oceanic to continental areas as shown in Figure 41. Under the oceans, the depth to the Moho may be as little as 8 km, whereas under the continents it averages 30 km, and may reach as much as 90 km beneath high mountain ranges such as the Andes and Himalayas. It seems that the configuration of the Moho mirrors the topography of the Earth's surface; the reason for this will be explained in Unit 6.

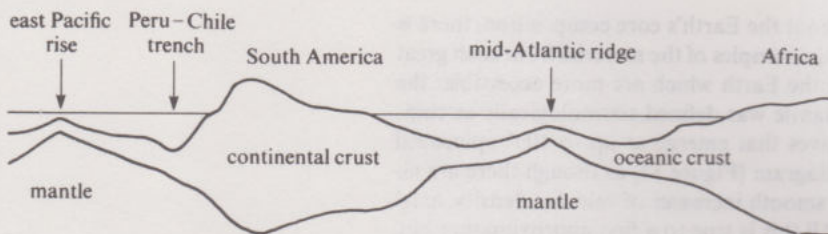


FIGURE 41 The thickness of the crust beneath ocean and continent shown schematically: the lower black line represents the Mohorovičić discontinuity.

Broadly speaking, the *continental crust* is composed of low-density igneous rock between *granite* (specimen S1) and *basalt* (S3) in average composition. Although granites are most important by volume in the upper parts of the continental crust, many *sedimentary rocks* cover them as a relatively thin veneer (a few km at most) at the surface. The rocks of the *oceanic crust* are rather less varied, consisting primarily of *basalt* (specimen S3). Ocean crust materials are slightly more dense than those of the continental crust because they contain rather more of the dense element, iron, and less of the lighter silicon. You will be taking a much more detailed look at the crust, its composition and the processes that take place therein when you reach Unit 27.

continental crust

ocean crust

4.6.2 The composition of the mantle

Apart from a very few special areas, the crust-mantle boundary lies at least several kilometres below the Earth's surface and so our sampling of the mantle is pretty restricted. Indeed, most of the mantle, extending up to 2900 km depth, is *totally inaccessible*. Once again, seismic information provides some useful clues about structure and composition; for example, we know that P-wave velocities increase suddenly from about 7 to 8 km sec⁻¹ at the Moho. There must be a marked difference in the physical properties of the materials above and below this boundary.

Earlier, we told you that the mantle is thought to be peridotite, a rock comprising mainly iron and magnesium silicates; what evidence do we have and does a peridotite mantle account for the observed velocity increase at the Moho?

First, we can determine experimentally that the density of peridotite is $3.3 \times 10^3 \text{ kg m}^{-3}$ and that its axial modulus is $2.1 \times 10^{11} \text{ N m}^{-2}$. Combined, these values give us a P-wave velocity of 8.0 km s⁻¹, just that which is found beneath the Moho.

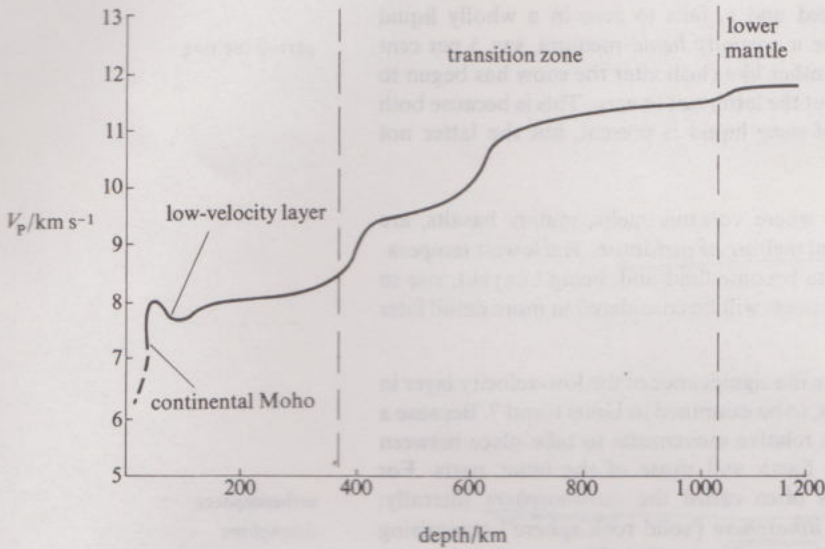
Second, fragments of peridotite are sometimes found in lavas erupted by volcanoes, and these lavas are believed to originate in the mantle at depths of 60 km or more (see Section 4.6.3). The peridotite may have been ripped off the walls of the volcanic conduit during eruption.

Third, there are a few places in the world (such as in Cyprus or in N. Italy, where specimen S4 was collected) where bits of mantle have been brought to the surface and peridotite is the predominant rock type.

It may seem that we have a watertight case for peridotite but, strong though the evidence is, *no one has ever seen the mantle in place*. Even the outcrops of mantle rock just mentioned are probably atypical, because they only represent the top few kilometres of a layer 2900 km thick! However, it is widely accepted that the Moho represents a change in physical and chemical properties of the materials—from granite or basalt above to peridotite below. The *chemical change* that occurs at the Moho is a change in the proportions of the elements present in these

various rocks and it is also the reason for the physical changes observed. The exact nature of the chemical change is another subject, which must be deferred to Units 27 and 28.

But is the entire mantle composed of peridotite? So far (Figures 34 and 35) we have implied that the entire mantle is homogeneous chemically and seismically. However, Figure 42 shows that there are some short steps in the seismic-velocity profile of the mantle between 400 and 1050 km depth. This region is called the *transition zone*, partly because its base separates the *upper* and *lower* mantle. Why is it transitional?



transition zone upper mantle
lower mantle

FIGURE 42 Simplified illustration of P-wave velocities in the Earth's upper mantle. This velocity-depth diagram has been calculated from detailed travel-time curves such as those in Figure 34.

Well, experiments on peridotite show that it undergoes certain changes at high states of compression equivalent to these depths. What would happen for example, if you squeezed a material until it was unable to withstand the pressure being exerted without changing the internal arrangement of its constituent particles? Such compression might cause a new, more dense and less compressible form of the *same chemical composition* to be produced. This is a *phase change**; a well-known example of a phase change between two solids is the change undergone by carbon from graphite (density $2 \times 10^3 \text{ kg m}^{-3}$) to diamond (density $3.5 \times 10^3 \text{ kg m}^{-3}$) at high pressures.

phase changes

So it is with peridotite in the upper mantle: there are *three separate phase changes* within the transition zone, where axial modulus increases rapidly but density also increases (see also Table 2). We think that the lower mantle contains iron and magnesium silicates in peridotite proportions: no phase or chemical change occurs and the region is characterized by a smooth increase of v_p with depth due to increasing compression.

low-velocity layer

4.6.3 The low-velocity layer

There is one other feature of the v_p variations in the upper mantle that we have not yet considered—see Figure 42. In 1926, the seismologist Gutenberg noted that seismic waves from earthquakes with foci located at depths between 50 and 250 km in the mantle took *longer* to arrive than they should have done according to simple calculations based on a homogeneous mantle.

What does this suggest to you about the material between 50 and 250 km?

Since the travel times are long, v_p and v_s must be lower. The layer has therefore come to be known as the *low-velocity layer*** (see Figure 42).

What would happen to a seismic wave from an earthquake whose focus is located within the low-velocity layer, which travels downwards into higher-velocity material?

* The changes from solid ice → liquid water → steam are phase changes between different states of the same chemical composition.

** Its presence has now been defined precisely by its effect on surface (L) rather than body waves, but the details do not concern us here.

Of course, it becomes refracted away from the perpendicular to the interface and when the critical angle is reached, total reflection will take place—see Figure 24 and Section 3.3.2.

Thus, some of the seismic waves became trapped within the low-velocity layer and they may never reach the surface because they follow a zig-zag path (Figure 24).

Can you think of one reason why seismic wave velocities might be lower in the low-velocity layer.

Remember that when we discussed the properties of the outer core in Section 4.5.1 we said that v_p is reduced and v_s falls to zero in a wholly liquid medium. Now, what if we have a *partially liquid* medium, say 5 per cent liquid and 95 per cent solid? (rather like slush after the snow has begun to melt). Both v_p and v_s will fall, but the latter not to zero. This is because both axial and rigidity moduli fall if *some* liquid is present, but the latter not to zero.

partial melting

The low-velocity layer is the region where volcanic melts, mainly basalts, are collected in the upper mantle by partial melting of peridotite. The lowest temperature melting components of peridotite become fluid and, being buoyant, rise to form *basalt volcanoes*. This melting process will be considered in more detail later in the Course.

Finally, it is important that you realize the significance of the low-velocity layer in terms of the Earth's physical processes, to be examined in Units 6 and 7. Because a partial melt exists, this region allows relative movements to take place between rocks of the outermost part of the Earth and those of the inner parts. For this reason the low-velocity layer is often called the *asthenosphere* (literally: weak sphere) and is overlain by the *lithosphere* ('solid rock sphere') comprising both the crust and uppermost mantle (see Figure 43a). The mantle below the low-velocity layer is called the *mesosphere* ('centre sphere'). You will find that these terms are used when dynamic physical processes within the Earth are being described, otherwise we use the traditional names based on seismic discontinuities.

asthenosphere

lithosphere

mesosphere

4.7 Summary of the Earth's structure and composition

Right at the beginning of this Unit, we suggested that you took careful note of how the structure of the *inaccessible interior* of the Earth has been assembled. We now have a pretty complete *model* based primarily on one set of measurements: the times taken for seismic waves to travel from an earthquake to various recording stations. Such fundamental observations allowed us to deduce the physical properties *density*, *axial and rigidity modulus* at depth in the Earth. From this we progressed, using independent evidence, to the view that the Earth has an iron-rich core that is partly liquid and partly solid. Other observations of seismic waves enabled us to recognize a boundary between the crust and the mantle (the *Mohorovičić discontinuity*), the existence of a *transition zone*, and a *low-velocity layer* in the mantle. The physical properties of the mantle were combined with surface observations and experimental data to deduce that the most appropriate mantle composition is a mixture of iron and magnesium silicates represented by the rock *peridotite*. The final structure we have deduced appears in Figure 43b which you should be able to reproduce. Some important properties of the different layers in the Earth are summarized in Table 2, which is intended to be a store of useful data for future reference.

Apart from learning about the Earth itself, you should also have gained some impression about how scientific models are constructed (cf. Unit 1, Section 1). In this case, starting with the simplest possible 'model'—a totally homogeneous Earth (Figure 33)—we introduced refinements, step by step, to yield our final, quite complex version (Figure 43b). The model seems to be a good one but, and we emphasize this point, *it is only a model*. There are major parts which are speculative and which may always remain so. For example, peridotite is the best candidate for the upper mantle, but no one has ever seen it *in place*—nor is anyone likely to. On the seismic information that Earth scientists have at present, the upper mantle might just as well be green cheese, as long as it is green cheese with a density of about $3.3 \times 10^3 \text{ kg m}^{-3}$ and the correct elastic moduli!

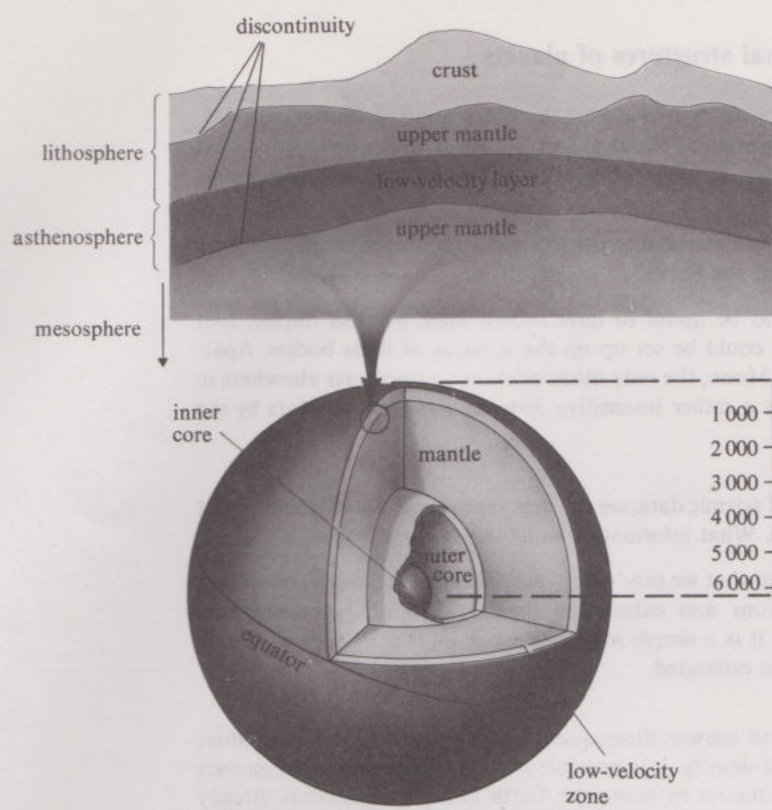


FIGURE 43 (a) Enlarged cross-section through the outer 400 km of the Earth with structure based on seismic data. (b) Cut-away section of the whole Earth showing its main subdivision as inferred from seismic data.

TABLE 2 Properties of the Earth's layers

Zones and discontinuities	Depth to boundaries km	Velocity of P-waves km s^{-1}	Percentage of mass	Density 10^3 kg m^{-3}	Axial modulus 10^{11} N m^{-2}	Possible nature of regions	Possible nature of boundaries
CRUST	continental: 25–90 (average 33) oceanic: 8–11	6.0 7.0	0.7	2.7 3.0	1.0 1.5	heterogeneous solids CONTINENTS: granitic OCEANS: basaltic	{ CHEMICAL CHANGE from basalt or granite to peridotite
Mohorovičić							
UPPER		8.0		3.3	2.1	various types of peridotite	{ PHASE CHANGE solid $\rightarrow$ solid + liquid $\rightarrow$ solid
Low-velocity layer	c. 50–250	7.8				partly molten peridotite	
MANTLE	400	8.2	68.0	3.4	2.3	solid peridotite	{ PHASE CHANGE to higher density minerals
Transition zone	1 050						
LOWER MANTLE		11.1		4.3	5.3	high-density minerals with overall peridotite composition	{ CHEMICAL CHANGE from silicate mantle to iron-rich core
core-mantle	2 900	13.6		5.4	10.0		
OUTER CORE		8.1		9.9	6.5	probably iron sulphide—liquid	{ CHEMICAL AND PHASE CHANGE
	5 155	10.1	31.3	12.3	12.5		
INNER CORE		11.2		c. 13.5	16.9	iron-nickel alloy—solid	
	6 370	11.3					

4.8 Postscript: the internal structures of planets

Units 1–3 started this Course by considering the Earth as an astronomical body. Unit 4 has focused on the interior of our planet. However, in the broader context of S101, let us finish by making reference to other bodies within our solar system.

Could you set about determining the internal structure of other planets in the same way as for the Earth?

Of course, it would be useful to have *seismic data*, but this implies that recording stations could be set up on the surfaces of these bodies. Apart from those on the Moon, the only other working seismometer elsewhere in the solar system is a rather insensitive instrument set up on Mars by the Viking missions.

Despite the lack of seismic data, we *can* make some useful statements about planetary interiors. What information must this be based on?

In Unit 3 you learnt that we can ‘weigh’ other celestial bodies by observing their orbital motions and calculating their gravitational accelerations. Having done this, it is a simple matter to work out the *density*, assuming that volume can be estimated.

We hope you were able to answer these questions. As you should now realize, once we have determined density, it is possible to make some intelligent guesses about other planetary interiors by using the Earth and Moon models already deduced from seismic data. What we actually find is that the ‘inner’ planets (Mercury, Venus, Earth and Mars) are relatively small and dense: they are probably all similar but with slightly different core-to-mantle ratios. Four of the five ‘outer’ planets (Jupiter, Saturn, Uranus and Neptune) are large planets with much smaller densities: by and large they are composed of light, gaseous rather than rocky material. The outermost planet, Pluto, is relatively small and so may be like the four inner planets. Unfortunately, it may be several centuries until we obtain good seismic data for the rest of the solar system but, once we do, there can be no doubt that the analysis of planetary internal structures will follow the same model-building technique that we have just used for the Earth.

Now that you have completed Section 4 of this Unit, you should be able to achieve the following additional Objectives:

- Indicate that you understand the meaning of the terms: *epicentral angle*, *surface waves*, *body waves*, *shadow zone*, *core–mantle boundary*, *outer core*, *inner core*, *iron meteorites*, *Mohorovičić discontinuity*, *continental crust*, *oceanic crust*, *phase change*, *transition zone*, *upper mantle*, *lower mantle*, *low-velocity layer*, *partial melting*, *lithosphere*, *asthenosphere*, *mesosphere*. Since there are rather a lot of new terms here, we have italicized the most important ones. (SAQ 18)
- Given data on the travel times of P-waves and S-waves for various distances from an earthquake epicentre, and given data on the average density of a planet and the average density of rocks found on its surface, draw correct conclusions about the variation of *elastic moduli* and *density* with depth below the surface. (ITQs 6 and 8, SAQ 19)
- Given data on the reception or otherwise—at various epicentral angles between 0° and 180° —of seismic waves transmitted through the interior of a planet, devise a plausible model for its *internal structure* including all *seismic discontinuities*. (ITQ 7, SAQ 19)
- Given a model for the internal structure and composition of a planet, predict qualitatively the existence and extent of P-wave and S-wave shadow zones. (SAQ 20)
- Draw a rough cross-sectional diagram of the Earth’s interior showing the principal components. (Figure 43b)
- Describe, in qualitative terms, how seismological and other data have led to certain conclusions about the principal rock types that are believed to be present at different depths within the Earth. (ITQ 9, SAQ 21)
- Explain qualitatively the importance of changes in phase and chemical composition in the origin of discontinuities in the Earth. (SAQ 22)

SAQ 18 Fill in each of the blank spaces (A–R) in the following paragraph using the number (1–17) representing the most appropriate term from the list provided at its end.

The travel-time curves for P-waves and S-waves in the Earth express the time taken for (A) to reach recording stations at various (B) around the surface. Such curves reveal the existence of a P-wave (C) , between 103° and 142° epicentral angles, which is caused by strong refractions at the (D) . Although this is the most important seismic discontinuity in the Earth, several others have been recorded and starting from the Earth's surface we have the (E) , which separates the basaltic (F) and more granitic (G) from the (H) , which is probably composed of (I) . This boundary represents a chemical change, but a little deeper in the Earth, temperatures are sufficiently high for a (J) of peridotite to form and between 50 and 250 km depth we observe a corresponding (K) . This part of the Earth's interior is sometimes called the (L) , in contrast to the relatively rigid, overlying (M) that comprises the uppermost mantle and crust. Both the low-velocity layer and the rather deeper transition zone are regions where the material of the mantle undergoes (N) , in the latter case to high-density minerals which characterize the seismically uniform (P) . Beneath the core-mantle discontinuity at 2900 km depth, we deduce that there is an iron-nickel rich (Q) with an iron sulphide (R) which together form about 31 per cent of the Earth's mass.

List of terms 1 epicentral angles; 2 body waves; 3 shadow zone; 4 core-mantle boundary; 5 outer core; 6 inner core; 7 Mohorovičić discontinuity; 8 continental crust; 9 oceanic crust; 10 phase changes; 11 upper mantle; 12 lower mantle; 13 low-velocity layer; 14 partial melt; 15 lithosphere; 16 asthenosphere; 17 peridotite.

SAQ 19 Figure 44 shows the travel-time curves for P-waves and S-waves in a remote hypothetical planet of similar size and average density to the Earth.

- Describe the variations in seismic wave velocities with depth in the planet and decide how many seismic discontinuities are recorded.
- Compare the possible variations in elastic moduli and density inside this planet with those in the Earth.
- In terms of the composition of the Earth's interior deduced in Section 4, decide upon possible compositions for the layers you have identified. How does the size of the layers compare with those in the Earth?

(Note that this is quite a time-consuming question, and we should be quite happy for you just to read the answer, especially if you are short of time.)

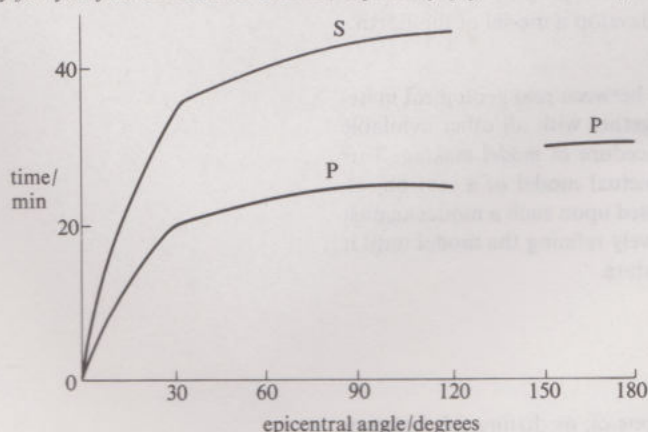


FIGURE 44 P-wave and S-wave travel times for a remote planet (SAQ 19).

SAQ 20 Seismic investigations are to be carried out on another remote planet whose interior is thought to contain only one seismic discontinuity separating the two layers: mantle and core. If the core occupies half the planetary diameter and v_p and v_s vary in the same way as in the Earth's lower mantle and outer core, predict the approximate extents of P-wave and S-wave shadow zones if (a) the core is liquid; (b) the core is solid and the ratio v_p/v_s is the same in the core and mantle.

SAQ 21 Select from the list below the reasons why the upper mantle is thought to be peridotite.

- (a) It has the right axial modulus, rigidity modulus and density.
- (b) It forms the floor of the deep ocean basins.
- (c) Fragments of peridotite are brought to the surface by volcanoes.
- (d) Volcanic islands are formed of peridotite.
- (e) Outcrops of possible upper mantle rock are peridotite.

SAQ 22 Without referring back to the earlier text, decide which of the following seismic discontinuities in the Earth are thought to be due to changes in phase and which to changes in composition:

- (a) Mohorovičić discontinuity
- (b) Low-velocity layer
- (c) Transition zone
- (d) Core-mantle boundary
- (e) Inner-outer core boundary

Aims and Objectives

Apart from Objective 1, which relates to all the terms and concepts used in this Unit, the Objectives may be divided into five groups which are related to the Aims of this Unit as follows:

Aims

- 1 (*Objective 2*) To describe the basic idea of the propagation of energy by waves such as water waves or seismic waves.
- 2 (*Objectives 3–8*) To explain what P-waves and S-waves are, and what determines their propagation velocities.
- 3 (*Objectives 9–15*) To explain how the laws of refraction and reflection follow from a simple basic idea (Fermat's principle) and how they may be applied to predict the paths of waves, such as seismic waves within the Earth, from hypotheses about the structure and wave velocities.
- 4 (*Objectives 16–22*) To show how measurements on the propagation of seismic waves through the Earth's interior can be used to develop a model of the Earth's internal structure and, eventually, composition.
- 5 (*Objectives 23–25*) To establish the relationship between real geological materials and their formative processes and use this together with all other available information to illustrate the general scientific procedure of *model making*. This involves starting with the simplest possible intellectual model of a real object, phenomenon or processes, checking predictions based upon such a model against observations and experimental data and progressively refining the model until it is capable of predicting satisfactorily all available data.

Objectives

- 1 Define correctly, or recognize the best definitions of, or distinguish between true and false statements concerning the terms and concepts listed in Table A. (SAQs 1, 4, 8, 11, 13, 14, 17 and 18)
- 2 Explain, by giving an example, that what is propagated by surface waves on water, or a seismic wave, or a wave down a row of balls and springs is not the medium itself but a state of disturbance and also energy. (SAQ 3)
- 3 Explain in simple terms why a shearing fracture of the rock of the Earth's crust can set up both P-waves and S-waves. (SAQ 5)

- 4 Explain in simple terms the variation in strength of P-waves and S-waves with respect to the direction of fault movement. (SAQ 5)
- 5 Explain how the velocity of a wave disturbance will vary when the stiffness or density of the propagating medium is changed. (ITQ 2, SAQ 3)
- 6 Outline the main steps in a semi-quantitative derivation of the formulae for the velocities of P-waves and S-waves through a solid medium. (This Objective relates to a non-essential text section.) (SAQ 9)
- 7 Explain in simple terms why an S-wave cannot be propagated through a liquid. (SAQ 10)
- 8 Calculate the velocities of P-waves and S-waves through a solid medium given the density and the relevant elastic moduli. (ITQs 3 and 4)
- 9 State Fermat's principle of least time.
- 10 State Snell's law of refraction and apply it to a wave passing through two media with different wave velocities. (SAQs 12 and 13)
- 11 Describe a graphical method for deriving Snell's law of refraction and apply it to a wave passing through two media with different wave velocities. (This is another non-essential Objective.) (ITQ 5)
- 12 As an extension of Objectives 8 and 10, given the relevant properties of two adjacent media, calculate the angle of refraction for any given angle of incidence and vice versa. (SAQ 15)
- 13 State the law of reflection. (SAQ 13)
- 14 Given the velocities of a wave in two adjacent media, predict the conditions under which total internal reflection will take place at the interface. (SAQ 14)
- 15 Explain qualitatively why a seismic wave should follow a curved path if the propagation velocity increases smoothly with depth below the Earth's surface. (SAQ 16)
- 16 Give a simple explanation for the difference between the average density of the Earth and the average density of rocks found at the Earth's surface. (ITQ 1, SAQ 2)
- 17 Distinguish, on a map of Earth, an approximately correct distribution of seismic zones from a grossly incorrect one. (SAQ 6)
- 18 Describe the principles of a simple seismometer. (SAQ 7)
- 19 Given data on the travel-times of P-waves and S-waves for various distances from an earthquake epicentre, and given data on the average density of a planet and the average density of rocks found on its surface, draw correct conclusions about the variation of *elastic moduli* and *density* with depth below the surface. (ITQs 6 and 8, SAQ 19)
- 20 Given data on the reception or otherwise at various epicentral angles between 0° and 180° of seismic waves transmitted through the interior of a planet, devise a plausible model for its *internal structure* including all *seismic discontinuities*. (ITQ 7, SAQ 19)
- 21 Given a model for the internal structure and composition of a planet, predict qualitatively the existence and extent of P-wave and S-wave shadow zones. (SAQ 20)
- 22 Draw a rough cross-sectional diagram of the Earth's interior showing the principal components. (Figure 43b)
- 23 Explain how the textures and grain size of different rocks provides useful information about the processes involved in their formation (SAQ 17)
- 24 Describe, in qualitative terms, how seismological and other data have led to certain conclusions about the principal rock types which are believed to be present at different depths within the Earth. (ITQ 9, SAQ 21)
- 25 Explain qualitatively the importance of changes in phase and chemical composition in the origin of discontinuities in the Earth. (SAQ 22)

ITQ answers and comments

ITQ 1 The density of the whole Earth, calculated in Unit 3, is about $5\,500\text{ kg m}^{-3}$, whereas we hope you obtained values between $2\,500\text{ kg m}^{-3}$ and $3\,300\text{ kg m}^{-3}$ for the four rock samples*. This leads to a reiteration of the conclusion made in passing during Section 6 of Unit 3 that the interior of the Earth must have a density greater than the average value.

By the way, the reason we asked you to measure 4 rock samples will become obvious in Section 4 of this Unit when we compare their individual density values.

* Our values were: S1 $2\,600\text{ kg m}^{-3}$ S3 $2\,900\text{ kg m}^{-3}$
S4 $3\,300\text{ kg m}^{-3}$ S6 $2\,500\text{ kg m}^{-3}$

ITQ 2 You probably thought that the engine might have to give the end truck a rather bigger nudge, a few feet perhaps, if any effect is to be seen at the far end of the train. True, but *this* train would resist being compressed much more weakly, and the push would be transferred relatively *slowly* along the train compared to one with rigid couplings. If the trucks were heavily loaded, the push would be transferred still more slowly. The important point is that to increase the propagation velocity of the 'pulse' you must decrease the mass or increase the rigidity of the couplings. In either case, the amount of force *between the trucks* is greater for a given amount of displacement, of course.

ITQ 3

$$v_p = \sqrt{\frac{\psi}{\rho}} = \sqrt{\frac{3 \times 10^{11}}{5.5 \times 10^3}} = 7\,385\text{ m s}^{-1}$$

$$v_s = \sqrt{\frac{\mu}{\rho}} = \sqrt{\frac{8 \times 10^{10}}{5.5 \times 10^3}} = 3\,814\text{ m s}^{-1}$$

Hence the velocity of seismic waves in a medium with these values of elastic moduli and density would be:

$$v_p = 7.4\text{ km s}^{-1} \quad v_s = 3.8\text{ km s}^{-1}$$

MAFS 1 (rounded off to two significant figures)

For $\sqrt{\frac{3 \times 10^{11}}{5.5 \times 10^3}}$, key: 3, EE, 11, $\div$, 5.5, EE, 3, =, $\sqrt{x}$, =

This gives you an impression of the rate at which seismic waves travel through the Earth ($v_p = 16\,650\text{ m.p.h.}$ and $v_s = 8\,500\text{ m.p.h.}$ in old-fashioned terms) an order of magnitude faster than Concorde!

ITQ 4 If you recall from the Home Experiment that density increases in the order air, wood, water, metal pipe, then the relationship:

$$\text{velocity} = \sqrt{\frac{\text{modulus}}{\text{density}}}$$

tells you that, if modulus is constant, velocity should *decrease* when density *increases*. So, to explain the *increase* in velocity for these materials, we deduce that *modulus is increasing faster than density*. You will find in Section 4 of this Unit that the same is true for the materials of the Earth's interior.

ITQ 5 Quite obviously the route shown in Figure 19 doesn't make sense, because you have to cover more distance at slower speed in the water than on land. So better routes must be like those shown in Figure 45 as a-f. We calculated the time taken for routes a-f and, if you measured distances on the land and sea with a sufficiently large piece of graph paper, you should have results, broadly in agreement with Table 3:

TABLE 3

Path	Running time/ seconds	Swimming time/ seconds	Total time/ seconds
a	17.7	35.4	53.1
b	19.5	32.0	51.5
c	21.5	29.2	50.7
d	23.6	26.9	50.5
e	25.6	25.5	51.1
f	28.0	25.0	53.0

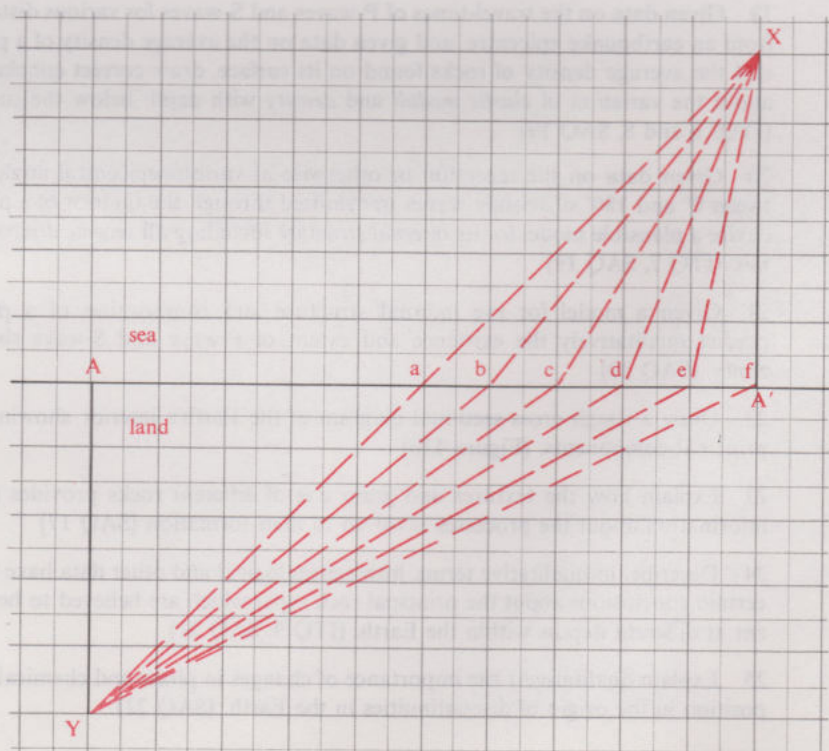


FIGURE 45 Possible routes from Y to X, taking times as calculated in Table 3.

Then we plotted a graph of the total time against the distance along the edge where we entered the water (Figure 46).

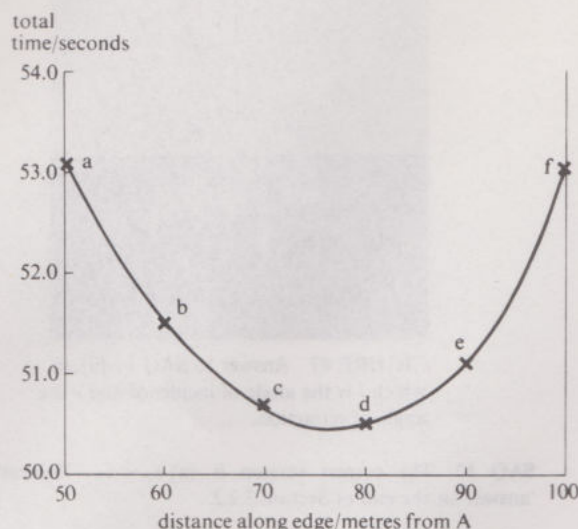


FIGURE 46 A time-distance graph showing the water entry point for least time.

As you can see from the graph, the path with the shortest total time must cross the edge somewhere between 70 and 80 metres from A—about half-way between c and d in fact. So we drew this particular path on our diagram and measured the angles θ_1 and θ_2 . The

values we found were $\theta_1 = 56^\circ$ and $\theta_2 = 24.5^\circ$. How did yours compare?

ITQ 6 An epicentral angle of 180° means that we have crossed one Earth diameter, which will take P-waves 37 minutes, 55 seconds, and S-waves 64 minutes, 20 seconds.

ITQ 7 The data indicate that the core model must apply and, moreover, that any shells of different density within the Earth must be spherically symmetrical in order that travel-time curves for different earthquakes should look identical. So an earthquake wave starting *anywhere* will reach a station *anywhere* around the circle located at a particular epicentral angle at the same time. This shows that although the Earth's *surface features* do not have spherical symmetry, the Earth's *interior* does. If the internal distribution of high-density material were random, as in the raisin model, then different geographic directions to recording stations about an earthquake would generate different travel-time curves.

ITQ 8 We could postulate that either:

- (a) The innermost core has a different chemical composition. In this case, the new material would have to be of lower density and/or much higher axial modulus.
- (b) Perhaps the material becomes solid again and this alone would involve a considerably increased axial modulus.

ITQ 9 In this case, pure iron is too heavy and so we must add something *less dense* to the outer core. Nickel is no good as an additive because it is even more dense than iron.

SAQ answers and comments

SAQ 1

- | | |
|-------------------------|------------------------|
| A 3, seismic waves | F 1, compression pulse |
| B 6, P-waves | G 4, seismology |
| C 7, S-waves | H 5, seismometers |
| D 9, wave propagation | I 2, energy |
| E 8, thought experiment | |

SAQ 2 Yes they are. The average density can be calculated from the data as follows:

Suppose the total volume of the Earth is $V \text{ m}^3$, the average density of the Earth is $D \text{ kg m}^{-3}$, and the total mass of the Earth is $M \text{ kg}$, then

$$D = M/V \text{ kg m}^{-3}$$

or

$$M = DV \text{ kg}$$

This applies to each layer of the Earth separately, i.e.

mass of inner core = density of inner core $\times$ volume of inner core
 $= 14 \times 10^3 \text{ kg m}^{-3} \times 0.01 \text{ V m}^3$

$$= 0.14 \times 10^3 \text{ V kg} \quad (1\% = \frac{1}{100} = 0.01)$$

mass of outer core = $10.6 \times 10^3 \text{ kg m}^{-3} \times 0.15 \text{ V m}^3$
 $= 1.59 \times 10^3 \text{ V kg} \quad (15\% = \frac{15}{100} = 0.15)$

mass of mantle = $4.5 \times 10^3 \text{ kg m}^{-3} \times 0.83 \text{ V m}^3$
 $= 3.74 \times 10^3 \text{ V kg} \quad (83\% = \frac{83}{100} = 0.83)$

mass of crust = $2.8 \times 10^3 \text{ kg m}^{-3} \times 0.01 \text{ V m}^3$
 $= 0.03 \times 10^3 \text{ V kg}$

So the sum of the four masses is $5.5 \times 10^3 \text{ V kg}$, and, of course, is equal to the total mass of the Earth, M (i.e. $0.14 + 1.59 + 3.74 + 0.03 = 5.50$).

Thus $M = DV \text{ kg} = 5.50 \times 10^3 \text{ V kg}$

whence $D = 5.50 \times 10^3 \text{ kg m}^{-3}$

(If you used your calculator, then the 'sum of the masses' calculation can be done 'all in one' by working in the memory. Key as follows:

14, EE, 3, $\times$, 0.01, =, M, +,

10.6, EE, 3, $\times$, 0.15, =, M, +,

4.5, EE, 3, $\times$, 0.83, =, M, +,

2.8, EE, 3, $\times$, 0.01, =, M, +, M, =

You should get an answer of 5.493, because your calculator has worked with more significant figures than the answer given above.)

SAQ 3 (a) A medium of lower density has less mass per unit volume ($D = M/V$). From $F = ma$, a constant force produces a greater acceleration a when mass m is decreased and, since acceleration is the rate of change in velocity (Unit 3), the propagation velocity will increase as density decreases. You could also argue that smaller masses have less inertia and so will move more rapidly than larger masses.

(b) When the earthquake waves arrive at the surface they can move large objects. This motion is the result of the transfer of energy to the objects, energy which must therefore have been transmitted by the earthquake waves.

SAQ 4 (a) Answer, B. A shows a stretching (tensile) force and C shows a compressional force.

(b) Correct answers are A, B, D and E. C, of course, describes the *magnitude* of an earthquake and not its intensity, and the instrument that detects earthquakes in F is a *seismometer*.

SAQ 5 (a) (i) There will be no P-waves in any of the six axis directions shown although there will be in any other direction. Although P-waves will start out in the Y and $-Y$ directions, they will produce compression pulses on one side of the fault plane (XY) and decompression (expansion) pulses on the other side of the fault plane, and these effectively cancel out.

(ii) S-waves will be propagated only along the directions (Z, $-Z$) and not along the directions (X, $-X$ or Y, $-Y$).

(b) (i) Because movement takes place in the Y direction, P-waves will reach maxima in four directions in the XY plane at 45° to the X and Y axes.

(ii) S-waves will have maximum strength in the Z, $-Z$ directions.

By the way, don't worry if you had difficulty answering SAQ 5. We hope you have read our answers. Please refer back to Figures 8 and 9 together with their associated text if you need extra help.

SAQ 6 Only zone D. Compare with Figure 10, p. 15.

SAQ 7 The simple seismometer (Figure 11) incorporates a massive body because this provides a component with sufficient inertia to be unaffected by movement of the Earth's surface on which the instrument is mounted. The rest of the seismometer, being firmly fixed to the ground, moves relative to the massive body and it is this relative motion that is recorded as a seismogram. Newton's second law of motion is being used in that the acceleration of a body under the action of a given force is inversely proportional to its mass. It is also true that the seismometer is designed so as to reduce as much as possible the force on this heavy weight due to Earth movements.

SAQ 8 All the statements are true except (a) and (c). The elastic modulus measures the ratio stress/strain not just the stress, and the compressional stress is F/A , not A/F .

SAQ 9

- A density
- B $1/\rho$
- C $v_p \propto \psi$
- D $v_p \propto 1/\rho$
- E Check the dimensions of v_p and ψ/ρ
- F $[L][T]^{-1}$
- G $[L]^2[T]^{-2}$
- H If v_p is measured in ms^{-1} and ψ/ρ in $(\text{ms}^{-1})^2$, then $k = 1$
- I v_s
- J μ

SAQ 10 (a) No, you would not: the seismometer A, on the surface of the ice, would register both P-waves and S-waves, but the one in the water would register P-waves only. This is because S-waves cannot be transmitted through a liquid, because liquids have no rigidity—their rigidity modulus is zero.

(b) Water is much less incompressible than ice: you need less stress to cause the same amount of deformation (strain) in water. Consequently, the axial modulus for ice is greater than for water. Since P-wave velocity is proportional to the square root of axial modulus, waves will travel faster in ice than water, reaching A before B.

SAQ 11 (i) (b) shows refraction; (a) shows reflection, which we shall discuss in the next Section.

(ii) See Figure 47.

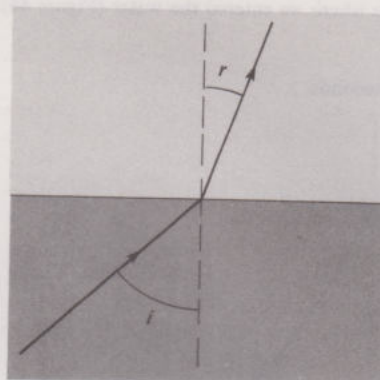


FIGURE 47 Answer to SAQ 11 (ii), in which i is the angle of incidence and r the angle of refraction.

SAQ 12 The correct version is (a) $v_A < v_B$. Remember the 'answer' at the end of Section 3.2.2.

SAQ 13 (a) angle of incidence: B
angle of reflection: A
angle of refraction: C

The angle D is none of these.

(b) law of reflection: $B = A$

$$\text{law of refraction: } \frac{\sin B}{\sin C} = \frac{v_1}{v_2}$$

SAQ 14 (a) Total reflection will be possible only if $v_A < v_B$.

(b) If $v_A = 0.5v_B$, the critical angle i_c is given by $\sin i_c = v_A/v_B = 0.5$; $i_c = 30^\circ$.

Hence, total reflection will occur for all angles of incidence from 30° to 90° .

SAQ 15 Answer: $r = 60^\circ$

Remember the formula for P-wave velocity is:

$$v_p = \sqrt{\frac{\psi}{\rho}} \quad \text{or} \quad (v_p)^2 = \frac{\psi}{\rho}$$

and this applies to both medium A and medium B, so that:

$$\frac{(v_p)_A^2}{(v_p)_B^2} = \frac{\psi_A/\rho_A}{\psi_B/\rho_B} = \frac{\psi_A \times \rho_B}{\psi_B \times \rho_A}$$

where ψ and ρ are the respective axial moduli and densities.

You also know that $\psi_A = 2/3\psi_B$

$$\text{and} \quad \rho_A = 2\rho_B$$

$$\text{So} \quad \frac{(v_p)_A^2}{(v_p)_B^2} = \frac{2}{3} \times \frac{1}{2} = \frac{1}{3}$$

$$\text{whence} \quad \frac{(v_p)_A}{(v_p)_B} = \frac{1}{\sqrt{3}}$$

$$\text{According to Snell's law, } \frac{(v_p)_A}{(v_p)_B} = \frac{\sin i}{\sin r} = \frac{1}{\sqrt{3}}$$

$$\text{So, if } i = 30^\circ, \sin i = 0.5, \text{ and } \frac{0.5}{\sin r} = \frac{1}{\sqrt{3}}$$

$$\text{or} \quad \sin r = 0.5 \times \sqrt{3} = 0.866, \quad \text{whence } r = 60^\circ$$

(On your calculator, the key sequence is: 0.5, $\times$, 3, $\sqrt{x}$, =, INV, sin)

SAQ 16 The wave path could show the changes of curvature illustrated only if velocity *decreased* with depth in layer A and then *increased* with depth in layer B.

SAQ 17 (a), (d), (e), (g) and (h) are true.

(b)—sedimentary rocks do not crystallize, they form by the cementation of pre-existing crystalline grains and fragments (see g).

(c)—granite and peridotite are *intrusive* rocks which crystallize *slowly* below the Earth's surface and so form *coarse* crystals. Only basalt is extrusive (see also d).

(f)—this is an entirely correct description of the products of *explosive* volcanism. Effusive volcanism is relatively quiescent and has more homogeneous basalt products.

SAQ 18

- | | |
|--------------------------------|--------------------------|
| A 2, body waves | J 14, partial melt |
| B 1, epicentral angles | K 13, low-velocity layer |
| C 3, shadow zone | L 16, asthenosphere |
| D 4, core-mantle boundary | M 15, lithosphere |
| E 7, Mohorovičić discontinuity | N 10, phase changes |
| F 9, oceanic crust | P 12, lower mantle |
| G 8, continental crust | Q 6, inner core |
| H 11, upper mantle | R 5, outer core |
| I 17, peridotite | |

SAQ 19 (a) There are three distinct velocity layers separated by discontinuities:

1 A low-velocity region for waves emerging up to 30° epicentral angles. Both P-waves and S-waves are present, the former travel about twice as fast as the latter. Both travel *much* more slowly than for similar waves in the Earth (see Figure 34), where, for example, P-waves reach 30° in 7 minutes compared with 20 minutes for this planet. Velocities of 2–3 km s⁻¹ for P-waves and 1–2 km s⁻¹ for S-waves must apply here.

2 Between 30° and 120° epicentral angles, velocities are much the same as in the Earth. P-waves take an additional 5 minutes for this part of their journey in both bodies at an average velocity of about 10 km s⁻¹.

3 A further discontinuity occurs at a depth equivalent to a 120° epicentral angle and beyond this there are P-wave and S-wave shadow zones. P-waves only re-emerge between 150° and 180° with somewhat reduced velocities. The waves take 31 minutes to traverse the straight line path to 180°, some 4 minutes longer than estimated by extrapolation of the layer 2 travel-time curve. This behaviour is similar to that on Earth (Figure 34) and, by analogy, P-wave velocities in the central region of this planet must be about 9 km s⁻¹.

(b) Layer 1 must have lower elastic moduli and/or higher density than the Earth's crust where P-waves travel twice as fast (remember $v_p = \sqrt{\psi/\rho}$ and similarly for S-waves. Layer 2 has higher values of v_p and v_s , not unlike those of the Earth's mantle and, in comparison with layer 1, ψ and μ must be higher and/or density lower. The former is more likely in view of the increase in *all three* physical quantities with depth in the Earth. Layer 3, with v_s zero and v_p

reduced, must be at least partly liquid (i.e. with zero rigidity modulus). This means that the axial modulus may also be reduced slightly but the fall in v_p also suggests that density is higher in this layer.

(c) To judge from the 30° epicentral angle of the first discontinuity, there must be several hundred kilometres of low-velocity crust. It is difficult for you to decide on a composition using only the information in this Unit, for v_p and v_s are lower than for the rocks we have considered. However, a high density is less likely than low values of elastic moduli, meaning that the rocks yield easily to applied stress. Although there are some terrestrial igneous rocks that have this property, it is a more common feature of sedimentary rocks.

Layers 2 and 3 have velocity characteristics similar to the Earth's mantle and core. In view of the similar average density for this planet, we suggest a solid peridotite layer 2 and a liquid iron-rich layer 3. You should note that this is a *tentative* conclusion based on even less evidence than for our Earth model!

We hope you also realized that the 120° discontinuity means that the mantle is larger and the core smaller than inside the Earth. Finally, there is *no evidence* in Figure 44 bearing on the possible existence of an *inner core*.

SAQ 20 Case (a) is like that of the Earth, except that the core is slightly smaller in relation to the planetary diameter. There will be shadow zones starting somewhere between 105° and 110° epicentral angles. No further S-waves will be recorded beyond this angle, but P-waves will re-emerge between about 145° and 180°. The angles 105° and 145° are estimations based on the fact that the core of this planet is half the diameter, whereas for the Earth the core is slightly greater than half the diameter, and angles of 103° and 142° result.

In case (b), S-waves will also re-emerge, and the size of the P-wave and S-wave shadow zones will be identical because Snell's law tells us that the amount of refraction depends on the *ratio* $v_{\text{mantle}}/v_{\text{core}}$, which is the same for P-waves and S-waves.

SAQ 21 You should indicate, first of all, that on the seismic evidence, peridotite has the right properties of rigidity, compressibility and density to fit the deduced seismic velocities at that depth; and on these grounds alone peridotite is perfectly acceptable—so you should have (a). We also know that fragments of peridotite are brought to the surface by volcanoes. It is suggested that pieces of the Earth's mantle could very easily be torn from the sides of a volcanic conduit by the ascending magma and be brought to the surface. This is indeed what we think has happened, so item (c) is further evidence. A third line of acceptable evidence is item (e), although we have no way of determining how anomalous are such outcrops of upper mantle.

Both the floor of the deep ocean basins and volcanic islands (b and d) are mainly basalt, though the former are often sediment covered.

SAQ 22 As you can verify by referring to Table 2 and the earlier text, discontinuities (b) and (c) are thought to be phase changes, whereas (a) and (d) are due to changes in composition. Discontinuity (e), at the inner-outer core boundary is actually a mixture of the two.

